

**URBAN SURFACE MODELS FOR BETTER AEROSOL RETRIEVAL WITH
MODIS AND LANDSAT**

by

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ABSTRACT

Urban surface models for better aerosol retrieval with MODIS and Landsat

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Aerosols are notoriously hard to measure on a global scale since they do not have unique spectral signatures like trace green house gases. Accurate global characterization of Aerosol Optical Depth (AOD) is essential because aerosols are the most uncertain mechanism in climate forecast models, and have known impact on human health. In particular, fine mode particulates (PM_{2.5}) can penetrate deep into the lung tissue contributing to lung damage and cardiac distress. Because of these effects on human health, the Environmental Protection Agency has strict monitoring standards for PM_{2.5}.

Aerosols measurements over urban areas are critical because extended urban centers can have significant aerosol loadings with air quality levels that are above EPA standards. For global studies, satellite measurements are the only realistic approach. Making this monitoring possible from space is the observation that column AOD is quite remarkably related to PM_{2.5}. Dark vegetative surfaces make such correlations strongest and more accurate aerosols retrieval. However, over urban scenes, it is particularly complicated due to the confusion between the ground signal and the aerosol signal. The

satellite sensors cannot distinguish if the incoming photons come from the surface or from atmosphere scattering.

For global retrieval of aerosols, the MODIS sensor is perhaps the most suited for global observations, because it can cover almost the entire planet in less than 2 days. The general approach is to use the Long Wave Channel (2130nm) as a good estimate of the surface albedo, since the aerosols contribution in this channel is almost always negligible (especially urban aerosols). Then, the surface albedos in the visible channels, where aerosols are important, can be inferred from empirical relations. However, it has become more apparent that the relations used by MODIS algorithms are not optimized for urban areas and tend to overestimate the AOD. This thesis provides a more extensive study of these relationships. By using high spatial resolution data, we analyzed the urban and vegetation areas separately to determine more accurate relations coefficients between the visible and 2130 nm channels to improve the MODIS algorithm for aerosols detection over urban areas. This work shows that for urban areas like New York City, higher values of these coefficients should be used. We showed that a much better estimation of AOD occurred when using these adjusted coefficients.

DEDICATIONS

This dissertation is dedicated to my beloved wife Maria who has had a very great deal of patience dealing with me throughout our years together while I tried to get this degree, and my daughters Yazel, Yamelli and Yailin for whom I am always thankful for bringing so much joy into our life.

I would also like to dedicate this thesis to my mother Austria for her undivided support and timely advice, and my father Adriano who first inspired me to seek higher education.

I also dedicate this to my family and friends who have stood by me and supported me.

And finally, to my older brother Adriano, a surgeon in the Dominican Republic, who looked out for me when we were kids but left my name out of his thesis (he claims that the reviewer did it). He continues to tell me that even with my Doctorate degree, he is still the only real Doctor in the family.

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CHAPTER 1

Introduction

The motivation for this thesis is the adverse effects of aerosols in humans, specially people with a compromised respiratory system and lung illnesses (like asthma), and the need of an accurate measurement of aerosols in highly populated areas. The main goal of this thesis was to find a more accurate method to measure aerosols over urban areas.

It is well known that accurate global characterization of Aerosol Optical Depth (AOD) is essential in quantifying energy balance for climate change studies [Charlson et al, 1992]. Aerosols can also have significant effects on human health. They pose a significant risk on human health [A-Saadi et al, 2005; Liu et al, 2007], with exposure linked to various health issues including heart and lung related distresses such as asthma. In particular, fine mode particulates (PM_{2.5}) are most important since these small particles penetrate deep into the lung tissue, which could lead to significant lung and cardiac distress.

Aerosols are notoriously hard to measure on a global scale since they do not have unique spectral signatures like trace green house gases such as CO₂ and CH₄. In particular, aerosol variability both in microphysical properties such as size and chemical composition and the diverse climate impacts make aerosols the most uncertain mechanism in climate forecast models. Making the situation even worse is the fact that

aerosols can further impact the climate via indirect processes. In particular, aerosols may affect both cloud albedo and precipitation mechanisms in complex ways making the magnitude of the impact very difficult to quantify.

For global studies, satellite measurements are the only realistic approach but aerosol retrievals over land are particularly complicated due to the confusion between the ground signal and the aerosol signal inherent in any passive solar reflectance measurement. Current approaches are usually associated with aerosol retrievals over Dense Dark Vegetation surfaces where the ground reflection is small. However, due to increasing land use and urban sprawl, retrieval of aerosols over urban areas are of increasing concern. One major reason that aerosol measurements over urban areas is critical is the obvious fact that extended urban centers such as New York City, Mexico City, Shanghai etc. can have significant aerosol loadings with air quality levels that are above EPA standards.

Making aerosols monitoring possible from space is the observation that column AOD is quite remarkably related to PM_{2.5}. Such correlations are strongest in the Eastern US where dark vegetative surfaces make the retrieval of aerosols more accurate. Unfortunately, other mechanisms make it very complicated to connect satellite derived AOD to surface particle mass. These include the variability of the Planetary Boundary Layer Height, the existence of aloft aerosol plumes as well as the natural variability of aerosols as manifested in the ratio of scattering to particle volume.

In retrieving aerosols over land, the approach used by the Moderate Resolution Imaging Spectroradiometer (MODIS) sensor is perhaps the one most suited for global observations. The general approach is to use the Long Wave Channel (2130nm) as a good

estimate of the surface albedo since aerosols are almost always negligible (especially urban aerosols). Then, the surface albedos in the VIS channels where aerosols are important can be inferred from empirical relations. However, it has become more apparent that the relations used by MODIS aerosols over land retrieval algorithms are not optimized for urban areas and a more extensive study of these relationships are needed.

The contents of this thesis are as follow: Later in this chapter, a brief motivation of the importance of aerosols for both human health and climate impacts are given. In chapter 2, we briefly discuss the different operational satellite algorithms that are being used to measure aerosols as well as the operational Collect 4 and Collect 5 aerosol retrieval algorithms used by MODIS. In chapter 3, we present the need to develop better spectral ratios between the visual (VIS) and Short Wave Infrared (SWIR) ground albedos and the results using long term MODIS data. In addition, the improvements in AOD using these updated values are given and the possibility of obtaining correlations using time series approaches is provided. In chapter 4, Hyperion sensor data processing is described and some of its consistencies and inconsistencies with MODIS are illustrated. In chapter 5, Landsat data processing is described and spectral coefficients for a number of observations are given, which are shown to be consistent with the previous results. In chapter 6, the Landsat spectral slopes are used to retrieve AOD at 300 meters resolution (10 x 10 pixels). In particular, consistency between neighboring vegetation and urban pixels in AOD retrieval can occur only if the modified spectral ratios are used. In chapter 7, conclusions and future work are given.

1.1 Aerosols definition and the importance to measure them

Aerosols are defined as minuscule solid and liquid particles suspended in the air. They range in size from 0.01 microns to several tens of microns. Aerosols may occur naturally, created by dust storms, sea spray, volcanoes activities, and fires from forests and grasslands due to lightning, for example, but they are also generated by human activities, such as the burning of fossil fuels and the alteration of natural surface cover, like clearing areas for urbanization. When averaged over the globe, aerosols contributions from human activities currently account for about 10 percent of the total amount of aerosols in our atmosphere [NASA Earth Observatory]. Aerosols are mostly found on the lower layer of the atmosphere, called the troposphere, but they can also be found on the stratosphere (right above the troposphere) [NASA Terra]. Under normal circumstances, the majority of aerosols form a thin haze in the troposphere, where they are washed out of the air by rain within about a week. Aerosols are an important but uncertain agent of climate change. They can alter atmospheric temperatures by absorbing and scattering radiation, thus depending on their type and location; they could warm or cool the troposphere. Aerosols can also modify clouds and affect precipitation. Sulfate aerosols can reduce cloud droplet size, making clouds brighter so that they reflect more solar energy, but black carbon aerosols strongly absorb solar radiation, warming the mid-troposphere and reducing cloud formation.

The following are two important reasons why do we care about aerosol and why are they so important to study and to find efficient ways to track and monitor them. Poor knowledge of the global distribution of aerosols is one of the reasons why there is a large uncertainty in climate prediction.

1.2 Effects of Aerosols on Global Climate

As previously stated, aerosols play a major role within Earth's climate system [IPCC, 2001], and affect the radiation budget, cloud processes, and surface air quality [Charlson et al. 1992]. With the exception of black carbon aerosols (which absorb solar radiation), aerosols tend to cause a cooling of the Earth's surface because most of them reflect sunlight back into space. Therefore they have a "direct" cooling effect by reducing the amount of solar radiation that reaches the surface. The magnitude of this cooling effect depends on the size and composition of the aerosol particles, as well as the reflective properties of the underlying surface. It is thought that aerosol cooling may partially offset expected global warming that is attributed to increases in the amount of carbon dioxide (a black carbon aerosol) from human activity.

Aerosols are also believed to have an "indirect" effect on climate by changing the properties of clouds (Figure 1.1). Aerosols form the nuclei for the water droplets that form a cloud. Indeed, if there were no aerosols in the atmosphere, there would be no clouds. It is very difficult to form cloud droplets without small aerosol particles acting as "seeds" to start the formation of cloud droplets. Water molecules attach themselves to aerosols to form drops. As the water drops grow larger, they eventually become large enough that they fall as raindrops. As aerosol concentration increases within a cloud, the water in the cloud gets spread over many more particles, each of which is correspondingly smaller. Smaller particles fall more slowly in the atmosphere and decrease the amount of rainfall. Because of this, changes in aerosols concentration in the atmosphere can change the frequency of cloud occurrence, cloud thickness, and rainfall amounts.

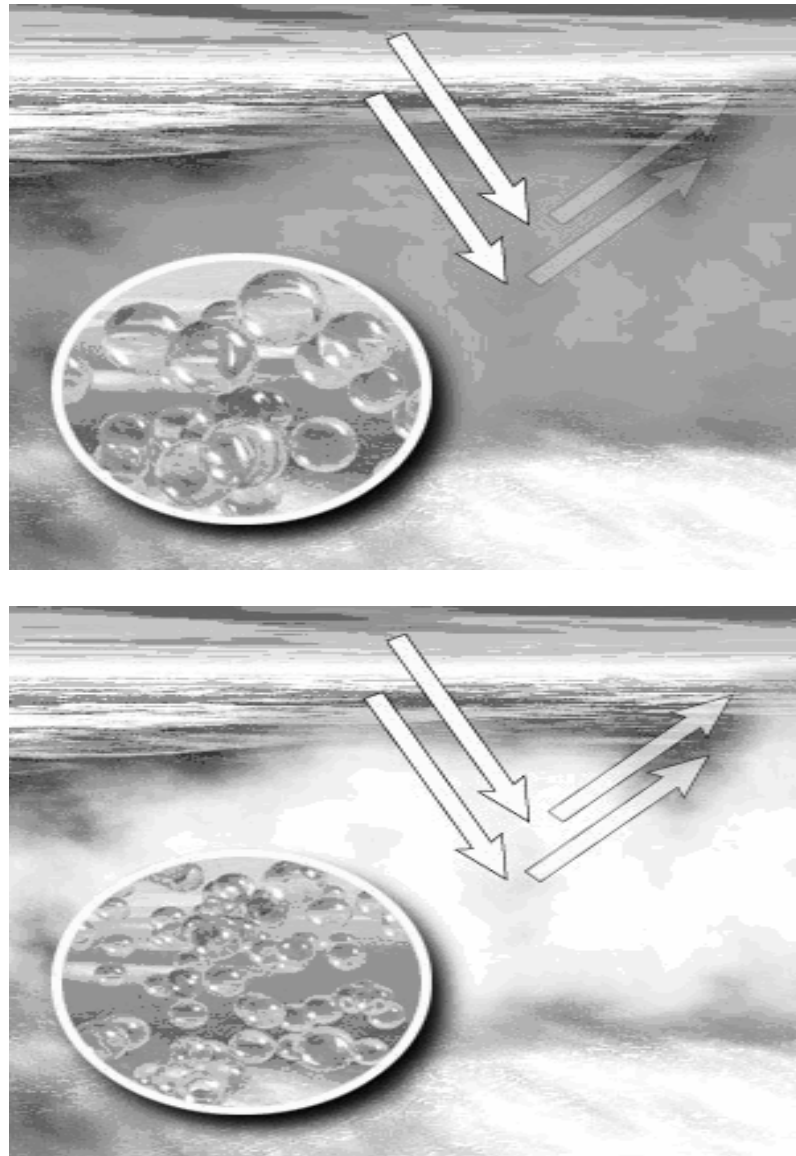


Figure 1.1: (Top) Clouds with low aerosol concentration have larger water droplets that do not scatter light well, and allow much of the Sun's light to pass through and reach the surface. (Bottom) The high aerosol concentrations in these clouds provide the nucleation points necessary for the formation of many small liquid water droplets. Up to 90% of visible light is reflected back to space by such clouds without reaching Earth's surface.

1.3 Effect of Aerosol on Human Health over Urban Area

Aerosols are also known as particulate matter (PM). The U.S. Environmental Protection Agency (U.S. EPA) divides aerosols in three categories: PM₁₀ for coarse aerosols (10 to 2.5 μm), PM_{2.5} for fine aerosols (0.1 to 2.5 μm), and PM_{0.1} for ultra fine aerosols (0.1 μm or smaller) [EPA website]. Research suggests that particle size is an important factor that influences how particles deposit in the respiratory tract and affect human health [A-Saadi et al, 2005; Liu et al, 2007]. Coarse particles are deposited almost exclusively in the nose and throat; whereas, fine and ultra fine particles generally are able to penetrate to deep areas of the lung. Fine and ultra fine particles are present in greater numbers and have greater surface area than larger particles of the same mass, and they are generally considered to be more toxic. Because these PM_{2.5} particles are so small, they are able to penetrate to the deepest parts of the lungs. Scientific studies have suggested links between fine particulate matter and numerous health problems. In particular, it has been well documented that adverse health effects from breathing air with a high PM_{2.5} concentration include premature death, increased respiratory symptoms and diseases, chronic bronchitis, and decreased lung function particularly for individuals with asthma [DEC]. In the United States, particulate matter is one of the six pollutants required monitoring as declared by the Clean Air Act (the others are Carbon Monoxide, Lead, Nitrogen Dioxide, Ozone and Sulfur Dioxide). The EPA monitors PM₁₀ and PM_{2.5} aerosols under the strict PM National Ambient Air Quality Standards (NAAQS), which sets 24-hour exposure at 35 $\mu\text{g}/\text{m}^3$.

The health effects of particulates matter are also heavily dependent on composition, its concentrations, and lifetime. In an effort to improve Air Quality

forecasting, the EPA has implemented an ambient air quality-monitoring program to determine the composition of airborne PM_{2.5} in urban air [AIRNow website], which produces air quality index maps as seen in figure 1.2, which shows the Northeast area. The yellow (light) areas on this map (which includes New York City) shows a moderate level of PM_{2.5}, which may cause problems for people sensitive to pollution.

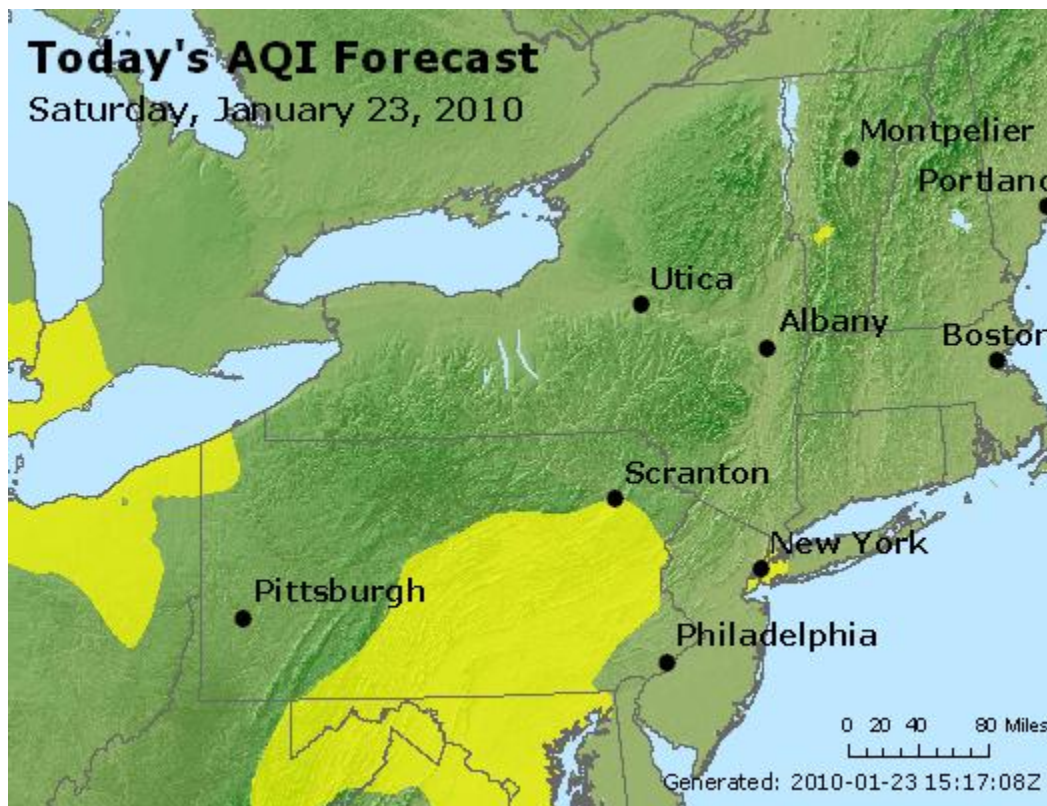


Figure 1.2 Air quality map (www.airnow.gov) for the Northeast. The yellow (lighter) areas (which includes New York City) shows moderate concentrations of PM_{2.5}.

There is also growing interest in using satellite retrievals of Aerosol Optical Depth (AOD) as a means to improve Air Quality Forecasts of PM_{2.5} and PM₁₀ measurements [Al-Saadi et al, 2005; Liu et al 2007; DEC website; Kittaka et al, 2004].

Systems such as IDEA (Infusing satellite Data into Environmental Applications) [Neil et al, 2004] routinely use MODIS AOD retrievals along with Lagrangian models to provide 24 hour predictions of PM_{2.5}. This system works by using static estimators which attempt to connect AOD remote sensing measurements directly to surface PM_{2.5} concentrations [Al-Saadi et al, 2005; Liu et al 2007; DEC website; Kittaka et al, 2004]. Further systems at research level are already endeavoring to assimilate AOD from satellites into air transport models [Kittaka et al, 2004].

In summary, the uncertainties created by the aerosols on global climate and their impact on human health makes it paramount to found reliable methods to measure them and track them.

CHAPTER 2

Current Satellite Approaches for Aerosols Measurement

Current aerosol over land algorithms have been applied to top of the atmosphere (TOA) radiance measured from satellites to retrieve aerosol properties, by separating the atmosphere and surface scattering contributions. Since different research team develops the different retrieval techniques, the characteristics of each algorithm are different in terms of spectral information, temporal and spatial resolution, and polarization availability. The characteristics of selected satellite instruments discussed in this paper are shown in Table 2.1. The algorithms for aerosol retrieval are often quite unique (sensor –specific) due to different characteristic of sensor, but all algorithms must account in some way in quantifying the land surface and obtain some estimate of aerosol type in order to retrieve aerosol optical depth.

Of course, all aerosol retrieval algorithms require accurate and robust cloud screening procedures. Cloud-contaminated pixels are not suitable for aerosol retrieval and unfortunately for comparisons, cloud masking procedures and techniques vary with the availability of wavelengths and view angles.

Sensor and Platform	AVHRR/3 (MetOp-A)	MISR (Terra)	AASTR (Envisat)	MODIS (Terra /Aqua)	POLDER (PARASOL)
Launch	Oct 2006	Dec 1999	Mar 2002	Dec 1999/ May 2002	Dec 2004
Resolution at nadir [km]	1.1	0.275	1.0	0.25 (bands 1-2) 0.50 (bands 3-7)	6 x 7
Swath width [km]	2399	2330	512	2330	2400
Multi view	No	Yes	Yes	No	Yes
Polarization	No	No	No	No	Yes

Table 2.1 General characteristics of current available aerosols over land sensors.

The most useful means of partitioning aerosol retrieval is how the different algorithms handle the land contamination. The three most widely used techniques are:

1) Estimates of ground reflection are made based on multiple time based observations of given pixels. In particular, it is hoped that for a large enough sampling, it is possible to locate “clear skies” in which the atmosphere signal is weak enough that reasonably accurate surface estimates can be obtained. Once these estimates are obtained, it is assumed that they are stable in time and can be used when aerosol loading is not small.

2) Simultaneous estimation of surface – aerosol parameters. In this case, using sophisticated optimization methods applied to multiangle, multispectral measurements over different times of observation are applied. In this scheme, the differences in time

periods when aerosols and surface properties can be expected to vary are directly incorporated.

3) Observation methods which are insensitive to ground but sensitive to aerosol signatures. This is clearly the most desired strategy but is hard to implement. At present, only polarization sensitive sensors such as POLDER and the soon to be launched APS on GLORY have this capability.

The following sections describe the techniques used by various sensors.

2.1 Single channel / single view polar sensors (AVHRR)

To begin, AVHRR is the simplest algorithm which illustrates the main features of aerosol retrieval. The AVHRR onboard POES is the earliest sensor design (see table 2.1) For AVHRR, only a single view angle and a single visible channel reflectance at $0.63 \mu m$ is available to retrieve aerosol optical depth (AOD). Since AVHRR does not generate operational land aerosol retrieval products, we will discuss research based algorithms.

Since AVHRR has a single channel, aerosol classification must be decided a priori based on historical climatologies. While such an approach is clearly less than optimum, it should be kept in mind that the long time series data makes the AVHRR processing very useful for long term climate (and environmental) data records (CDR or EDR).

In processing AVHRR data, the Pathfinder-Atmosphere (PATMOS) dataset is used [Knapp et al, 2002; Jacobowitz et al 1999; Stowe et al. 2002]. The PATMOS dataset is actually statistically calculated cloud-free radiance data for each day on a global 110

km grid for all five channels of AVHRR from 1981 through 2001. The retrieval of aerosol information from PATMOS data over land is a three-step process.

1. First, TOA reflectances for numerous conditions are calculated from the DISORT radiative transfer model version 2 and stored in look up table (LUT) [Stamnes and Swanson, 1981].
2. Second, these tables are then used to retrieve the surface BRDF from the PATMOS data. For example, over the course of 24 days it is possible to composite 3 observations from each look angle given cloud-free observations. The LUT with AOD equal to 0.05 is used to determine the surface BRDF. The LUT is interpolated for the viewing geometry of each of the points, and then retrieve the best-fit BRDF parameters.
3. Third, the surface BRDF data are then used to retrieve AOD from the PATMOS data using LUT information for that BRDF.

In a more sophisticated approach, [Hauser et al., 2005 a, b], no a priori climatological knowledge of the surface reflectance is necessary. Hauser's algorithm assume that the surface reflectance remains radiometrically invariant over a certain time period, which is 45 days in his algorithm, and some of the AVHRR observations are made under aerosol-free conditions. The following are main points in Hauser's algorithm.

1. Identify the aerosol-free data points using the concept of the convex hull, which is plotting the surface reflectance as a function of the satellite zenith angle and finding the minimum reflectances in terms of satellite angles.

2. Extract the surface reflectance (pixel-wise BRDF corrected surface reflectance) for background conditions for the whole range of satellite zenith angles from the aerosol-free data.
3. Retrieve aerosol optical depth using a radiative transfer model, after the surface reflectance is known and assumptions concerning the characteristics of aerosols are made.

The radiative transfer code used in Hauser's algorithm is the Simplified Method for Atmospheric Correction (SMAC) [Rahman and Dedieu, 1994], which is based on the 5S code [Tanre' et al., 1990] and 6S code [Vermote et al., 1997].

This procedure of course is severely limited by the fact that significant observations for estimating the surface BRDF for clear sky conditions is hard to implement since it is a complex function of the geometry. In particular, this procedure is clearly ill suited for urban aerosol retrieval. Later we will see that with multiple wavelengths, a significant improvement is possible since the spectral variation of surface angular dependence is weak.

2.2 Multi-angle without long-wave channel (AASTR and MISR)

Since AASTR and MISR are multi-angle imaging radiometer, both algorithms take advantage of differing angular reflectance signatures of the surface and atmosphere to accomplish retrieval of aerosol optical properties and optical depth over such surfaces.

2.2.1 AASTR dual-angles aerosol retrieval algorithm

The algorithm is based on the simple idea that surface models can be found which are very accurate radiometrically and can be parameterized in a very simple manner on 2 parameters: (1) a structural parameter v , which depends only on viewing angle and (2) the spectral parameter w that depends only on wavelength. The effect of atmospheres is then included by modifying the surface model to account for the degree of anisotropy of the downwelling radiation. Since scattering of light in atmosphere varies with size parameters of aerosol, which is $2\pi r_{\text{eff}}/\lambda$, and shorter wavelength has greater scattering ability. Therefore, the effect of atmospheric scattering on anisotropy surface varies with both the wavelength of the scattering light and the shape of the underlying surface BRDF. All these effect are taken into account in the BRDF model

$$\rho_{\text{surf}}(\lambda, \theta) = (1 - D(\lambda))v(\theta)w(\lambda) + \frac{\gamma w(\lambda)}{1 - g} [D(\lambda) + g(1 - D(\lambda))], \quad (2.1)$$

where $g=(1-\gamma)w(\lambda)$, λ is the wavelength, θ is the viewing angle (in AASTR, forward or nadir) ρ_{mod} is the modeled BRDF reflectance, γ is the fraction contribution to higher order scattering and set to 0.3 [North et al., 1999], and D is the fraction of diffuse irradiance. The first and second terms of (2.1) refer to direct and diffuse scattering, respectively. This model of surface scattering has been shown theoretically to a tractable inversion method, which is potentially more robust than the simple assumption of angular invariance alone [North et al., 1999; Grey et al., 2006]. This model can be used to estimate the degree of atmospheric contamination for a particular set of reflectance measurements and to find the atmospheric parameters that allow retrieval of a realistic surface reflectance [Grey et al., 2006].

In this algorithm, for a fixed AOD and aerosol model, the optimized surface parameters can be obtained that fit the multispectral and multiangle information the best and the residual can then be minimum keeping in mind that due to the good assumption that the angular surface model is not sensitive to angle, we only have 2 angular surface unknowns. Therefore, 8 TOA reflectances (4 wavelengths x 2 angle) are used to optimize the 4 wavelength factors of the surface and the 2 angle parameters for each AOD. A global minimization over the admissible AOD is then performed for the retrieval. In this model, it is necessary that assumptions are made to the other aerosol optical properties in aerosol models selection (smoke, urban, desert dust, continental and maritime) [Grey et al., 2006]. The selection of aerosol model depends on the location and time of the year. To retrieve AOD from dual view angle TOA reflectances ((550, 670, 870, 1630 nm) bands in the nadir and forward views), a coupled inversion (loosely described above) of the atmosphere path reflectance and the model of the surface scattering shown in equation (2.1) is applied.

Step by step inversion procedures are as follow [North et al, 1999; Grey et al, 2006]:

1. Retrieve a set of eight ground reflectance values (by using surface scattering model) and estimates of diffuse irradiance (D) at the four wavelengths given an initial estimate of the atmospheric aerosol model and AOD at 550 nm by inversion of the 6S radiative transfer model.
2. The best fit of the two structural parameters $v(\text{nadir})$ and $v(\text{forward})$ and four spectral parameters $w(555)$, $w(660)$, $w(870)$, and $w(1610)$ to surface scattering model equation are found by using the iterative minimization method of Powell.

The fraction of diffuse irradiance D is calculated in 6S from the AOD value and aerosol model used for that interaction.

3. Use Brent line minimization through a range of AOD to converge on the optimum value for AOD. For each iteration, each estimate of AOD results in a different set of surface reflectance values.
4. When the best fit of the parameters for a given AOD is found, while iteration is calculated using a different estimate of AOD. This process continues until the interaction between AOD and the two v and w parameters leads to the overall best-fit solution. The dual-view algorithm makes use of the full information content of observations and incorporates a set of eight reflectance values in the four AATSR optical channels and at the two looks.
5. The optimum value of AOD is the best fit of surface reflectances to equation (2.1) and is attained by minimizing the error function E_{mod}

$$E_{\text{mod}} = \sum_{\theta=1}^2 \sum_{\lambda=1}^4 [\rho_{\text{surf}}(\lambda, \theta) - \rho_{\text{mod}}(\lambda, \theta)]^2 \quad (2.2)$$

where ρ_{mod} is the surface reflectance estimated using equation (2.1) based on the best-fit values of the parameters v and w , and ρ_{surf} is the surface reflectance calculated using 6S given the TOA reflectance ρ_{toa} and the estimated atmospheric profile.

However, we again meet with the fundamental problem that for urban surfaces, the above atmosphere - surface coupling parameterization has not been tested and appears far to simple for bright surfaces where multiple interactions have to be handled with more care.

2.2.2 MISR multi-angles aerosol retrieval algorithm

MISR [Diner et al., 1998] is observing the Earth on the polar-orbiting Terra platform (on the same platform with the MODIS-Terra) at an altitude of 705 km. MISR measured the TOA radiance by using nine cameras, each fixed at a particular view zenith angle in the along-track direction and having four spectral bands (446, 558, 672 and 866 nm) with a cross-track ground spatial resolution of 275 m or 1.1 km. The fore-aft cameras are paired in a symmetrical arrangement and acquire multispectral images with nominal view angles relative to the Earth's surface of 0° , 26.1° , 45.6° , 60.0° and 70.5° , labeled An, Af/Aa, Bf/ Ba, Cf/Ca and Df/Da, respectively. The letters A–D denote the lens design, each lens system progressing from shortest focal length (A cameras) to longest (D cameras) to preserve cross-track spatial resolution, and the letters 'n', 'f' and 'a' indicate nadir, forward and aftward views, respectively.

The objective of MISR aerosol retrieval algorithm is using the advantages of multi bands and multi angles viewing capability to separate scattering radiance from atmosphere and underlying surface without *a priori* knowledge of the land surface type. From these two separate radiance components, determination the aerosol properties and the surface reflectance characteristics can be done. In current version of retrieval algorithm, two retrieval algorithms steps are included. The original version is based on the principal components analysis of multispectral and multi angle TOA MISR data from which empirical orthogonal functions (EOFs) for each wavelength for all camera angles were obtained to describe the directional reflectance properties of the surface [Martonchik et al., 1998a]. The upgraded version algorithm was implemented after 3

years of operation [Diner et al., 2005] by adding the aerosol retrieval algorithm based on the surface reflectance angular shape similarity already discussed with AASTR analysis.

The mathematical details of the latest MISR retrieval algorithm are quite involved and out of place in this thesis. However, it has been found [Diner et al, 2005] that MISR retrievals in urban areas are significantly underestimated. This together with the lack of spatial coverage makes MISR less suited as the best urban aerosol monitoring instrument. In the next section, the workhorse MODIS sensor is discussed in detail.

2.3 Multispectral Retrieval from MODIS.

The MODIS instrument is aboard the Terra (launched in late 1999) and Aqua (launched in early 2002) satellites as part of NASA's Earth Observing System (EOS) [NASA Terra; NASA Aqua]. These polar orbiting satellites orbit the Earth in opposite direction to provide global data acquisition. Terra (EOS AM) has a north to south orbit that is timed to cross the equator in the morning. Aqua (EOS PM) has a south to north orbit that is timed to cross the equator in the afternoon. Terra MODIS and Aqua MODIS can cover the entire Earth's surface every 1 or 2 days. The MODIS instrument acquires data in 36 spectral bands from 0.4 to 14.4 μm and spatial resolutions of 250 m, 500 m and 1 km. Bands 1 through 7 (see table 2.2) are primarily used in the studies of land, clouds and aerosols properties.

Band	Bandwidth (nm)	Spatial Resolution (m)	Spectral Radiance (W/(m ² -μm-sr))	SNR
1	620-670	250	21.8	128
2	841-876	250	24.7	201
3	459-479	500	35.3	243
4	545-565	500	29.0	228
5	1230-1250	500	5.4	74
6	1628-1652	500	7.3	275
7	2105-2155	500	1.0	110

Table2.2 MODIS specifications for bands 1-7 used for land, clouds, and aerosols properties analysis.

There are six Level-2 (Orbital Swath) MODIS Atmosphere products collected from Terra platform and is assigned an 8-character Earth Science Data Type (ESDT) name, which is used in cataloging and archiving the datasets. The Level-2 MODIS Atmosphere products are: (1) Aerosol Product, (2) Water Vapor Product, (3) Cloud Product, (4) Atmosphere Profile Product, (5) Cloud Mask Product, and (6) Post-launch Joint Atmosphere Product [Level-2 MODIS].

While channels 1 through 7 are used for aerosol retrieval over water, difficulties in obtaining estimates of the ground reflection make it impractical to use all of these channels for aerosol over land retrieval, therefore only a few of these channels have been dedicated to aerosol retrieval over land. In particular, the retrieval over land uses only

channels 1, 3 and 7 ($0.66\mu\text{m}$, $0.47\mu\text{m}$, and $2.12\mu\text{m}$ respectively) and uses assumptions for aerosol type based on geographical climatology [Remer et al] as well as assumptions on the surface albedo for the VIS channels. Based on optimal fitting of the appropriate top of the atmosphere (TOA) reflectance to appropriate radiative transfer Look Up Tables (LUT), the MODIS operational algorithm retrieves both the Aerosol Optical Depth (AOD or τ) and the mixture between fine and coarse mode (i.e. fine mode weighting FW or η).

2.3.1 MODIS Previous Operational Algorithm: Collection 4

As discussed earlier, it is best to work with dense vegetation targets. However, even moderately dark vegetation scenes have a residual reflectance that cannot be realistically ignored unless the aerosol optical depth is high. Therefore, some estimate of the ground reflectance is essential. To this end, it was noticed that the surface reflectance across the solar spectrum is correlated to some extent. Soils usually have an increasing reflectance as a function of the wavelength with correlation between the reflectances slowly decreasing with an increase of the wavelength span. Parallel physical processes affect the reflectance in the 0.47 and $0.66\mu\text{m}$ channels and in the 2.1 and $3.8\mu\text{m}$ channels. For example, the presence of vegetation decreases the reflectivity in the visible channels due to chlorophyll absorption and in the MIR (middle infrared) channels due to absorption by liquid water associated with the plant [Kaufman and Remer 1994]. In particular, wet soil has a lower reflectance in the visible channels due to light trapping, and in the 2.1 and $3.8\mu\text{m}$ channels due to the liquid water absorption.

To utilize this correlation, it is necessary to observe that in the 2.1 and $3.8\mu\text{m}$ channels, the aerosol signature is negligible and the TOA reflectance in the MIR can be

taken as the ground reflectance. Using correlations based on a-priori estimates of land cover type, the 0.47 and 0.66 μm ground reflectances can be estimated thereby allowing an improved estimate of atmospheric reflectance [Martins et al 2002]. Finally, as pointed out in [Kaufman et al 1997], the 3.8 μm channel is often contaminated by significant thermal emission and is therefore not used in the operational MODIS aerosol retrieval over land unless it is expected that large aerosol modes such as Saharan dust and/or smoke are present. However, these ground correlations would be useless if the 2.12 μm reflectance which we normally assume to be ground reflection has a significant aerosol contribution. Unlike the C004 algorithm, as we will discuss, the C005 algorithm is capable of isolating out the aerosol contribution.

Finally, it is necessary to point out that even with accurate atmospheric path reflectance values, aerosol retrieval is hardly straightforward. In order to derive the optical thickness from the atmospheric reflection, the aerosol size distribution, single scattering albedo and refractive index have to be assumed. Sensitivity studies showed that in a general case these assumptions generate substantial errors in the derived aerosol optical thickness ($\sim 30\%$). To reduce the errors, a good model of the aerosol properties based on measurements is required [Martins et al 2002; Remer et al 2001]. In regions where the model is most applicable, we can expect the remote sensing procedure to be more accurate.

The above approach has been shown to be useful over dark and fairly uniform land cover types [Wen et al 1999]. Unfortunately, these conditions are not realistically met over large urban areas such as the New York City. In fact, as pointed out in [Kaufman and Tanre 1998] and [Remer et al 2001], the retrieval of MODIS AOD over

the northeast coast of U.S. resulted in remarkably high overestimation when compared to sunphotometer derived data. It is quite reasonable that such overestimations was a result of poor land surface assumption. However, quite erroneously, the AOD overestimation results were attributed to high spatial inhomogeneity and sub pixel water contamination but no quantitative calculations were offered for this explanation. However, when the MODIS C004 aerosol product was validated by comparison with AERONET derived aerosol product, over the U.S. East coast, the MODIS retrievals tended to overestimate AOD (or τ) in clean conditions (i.e. low aerosol optical depth conditions), and underestimate in more hazy conditions[Remer et al; Chylek et al, 2005].

Additional weaknesses in the C004 approach included incomplete aerosol model capabilities, an assumption that the aerosol reflectance at 2.12 μ m is negligible which invalidates cases where coarse mode aerosols are significant and surface models most associated with vegetation. To address these issues, modifications were built into the most recent operational algorithm Collection 005 (or C005) [Levy, Remer et al 2005].

2.3.2 MODIS Latest Operational Algorithm: Collection 5

The MODIS aerosol retrieval operational algorithm over land is based on a LUT approach, i.e., radiative transfer calculations of TOA reflectance are pre-computed for a set of aerosol models (4 fine modes and 1 coarse mode). The MODIS land operational LUT contains pre-computed optical properties of aerosol at four discrete wavelengths (0.466, 0.553, 0.644 and 2.119 μ m, representing MODIS channels 3, 4, 1 and 7, respectively) for several values of aerosol total loadings, and for a variety of geometry. For discrete optical depths (described by the AOD at 0.55 μ m) each spherical aerosol

model (Continental, Generic, Absorbing smoke and Non-absorbing urban) and non-spherical model (Dust), scattering/extinction properties of aerosol size distributions are calculated by either MIEV [Wiscombe, 1981] or the Dubovik's T-matrix code [Dubovik et al., 2000, 2002a, 2002b]. Also LUT apply the assumption of a Rayleigh atmosphere and realistic layering of the aerosol by using US Standard Atmosphere, more detail can be seen at <http://modelweb.gsfc.nasa.gov/>.

Like many other passive sensor retrieval algorithms, the principle behind aerosol retrieval over land in MODIS is also based on the fact that the TOA reflectance is the combination of the atmospheric path reflectance and the ground reflectance and written as:

$$\rho_{\lambda}^{TOA}(\mu, \mu_0, \phi - \phi_0) = \rho_{\lambda}^{atm}(\mu, \mu_0, \phi - \phi_0) + \frac{T_d(\mu, \mu_0, \phi - \phi_0)T_u(\mu, \mu_0, \phi - \phi_0)\rho_{\lambda}^{surf}}{1 - S_{\lambda}\rho_{\lambda}^{surf}} \quad (2.3)$$

For the LUT, all the necessary atmospheric parameters in equation (2.3) such as upwelling and downwelling diffuse transmissions and aerosol spherical albedo, T_u , T_d , and S respectively, are calculated for seven aerosol loadings ($\tau_{0.55\mu m} = 0.0, 0.25, 0.5, 1.0, 2.0, 3.0$, and 5.0) with 9 solar zenith angles ($\mu\theta = 0.0, 6.0, 12.0, 24.0, 35.2, 48.0, 54.0, 60.0$ and 66.0), 16 sensor zenith angles ($\mu = 0.0$ to 66.0 with increments of 6.0), and 16 relative azimuth angles ($\phi_0 = 0.0$ to 180.0 with increments of 12.0).

Because of MODIS over-land retrieval employs only three channels, it is not able to retrieve the fine aerosol model [Remer et al., 2005]. Therefore, MODIS aerosol retrieval algorithm must assign the fine aerosol model a priori of the retrieval by using cluster analysis. In particular, the aerosol model is assigned as a function of season and location from a global aerosol models obtained from AERONET [Remer et al., 2005; Levy et al 2006a, 2006b].

The details of the aerosol retrieval in C005 are outlined as follows:

1) TOA reflectance signal (cloud cleared Level-2 product in 10 km resolution) at band 7 (2.12 μ m) measured by MODIS sensor can be converted to surface reflectance at the Short Wave Infrared (SWIR) by computations with pre-calculated LUT atmospheric parameters. The AOD is unknown at this stage; therefore surface reflectance at SWIR varies as a function of AOD (τ), (i.e.; AOD = 0, 0.25, 0.5, 1.0, 2.0, 3.0, and 5.0) for two size parameters (one fine mode and one coarse mode aerosol model). SWIR surface reflectance can be calculated as Eq. (2.4)

$$\rho_{2.12}^s(\tau, size) = \frac{\rho_{2.12}^{atm}(\tau, size) - \rho_{2.12}^{measured}}{s(\tau, size)(\rho_{2.12}^{atm}(\tau, size) - \rho_{2.12}^{measured}) - Td(\tau, size)Tu(\tau, size)} \quad (2.4)$$

2) Given the SWIR surface reflection (for each aerosol assumption), the visible surface reflectance can be calculated from the SWIR 2.12 μ m surface reflectance based on angular similarity assumptions used repeatedly in this chapter. In particular, based on global cluster analysis techniques, relationships connecting the visible and SWIR ground reflectances can be modeled based on two parameters, a Modified Vegetation Index (MVI on eq. 2.5), which provides a first estimate of surface classification, and the scattering angle Θ .

$$MVI = \frac{\rho_{1240}^{TOA} - \rho_{2120}^{TOA}}{\rho_{1240}^{TOA} + \rho_{2120}^{TOA}} \quad (2.5)$$

Based on sensitivity studies, these two parameters were found to be the most important in defining the VIS-SWIR reflection relationships. This is done by correlate the SWIR surface reflectance into visible surface reflectance directly. Note also that the MVI parameter is less sensitive to aerosols than the more used Normalized Difference Vegetation Index (NDVI) and is therefore more robust when aerosol loading is

significant. The new relations between the SWIR and the VIS channels surface albedos

are given as:

$$\begin{aligned}\rho_{0.66}^s &= f(\rho_{2.12}^s, \Theta, MVI) \\ \rho_{0.47}^s &= g(\rho_{0.66}^s, \Theta, MVI)\end{aligned}\quad (2.6)$$

Given the surface reflections at 0.47 μ m and 0.66 μ m, the TOA reflectance for all 3 bands (0.47 μ m, 0.66 μ m and 2.12 μ m) can be recomputed from atmospheric parameters within the LUT. To account for both fine and coarse modes, a self-consistent mixture parameter is retrieved in the process. To see how this is done, we write the individual path reflectance components for both fine and coarse modes:

$$\rho_{\lambda}^{f,c}(\tau) = \rho_{\lambda}^{af,c} + T_d^{f,c} T_u^{f,c} \rho_{\lambda}^s / (1 - s_{\lambda}^{f,c} \rho_{\lambda}^s) \quad (2.7)$$

where ρ_{λ}^{af} and ρ_{λ}^{ac} are the fine and coarse model atmospheric path reflectance, $T_{d\lambda}^f$ and $T_{d\lambda}^c$ are normalized downward fluxes for zero surface reflectance, T_{λ}^f and T_{λ}^c represent upward total transmission into the satellite field of view, and s_{λ}^f and s_{λ}^c are atmospheric albedo. Then, for a given ratio η of fine and coarse mode aerosol, the total reflectance can be written as the superposition:

$$\rho_{\lambda}^* = \eta \rho_{\lambda}^f + (1 - \eta) \rho_{\lambda}^c \quad (2.8)$$

where ρ_{λ}^f and ρ_{λ}^c are the each composite of the surface reflectance ρ_{λ}^s and atmospheric path reflectance of the separate aerosol model. Note that this superposition is possible since we are combining only components with the same optical depth so the normalized weighted superposition has the same optical depth as the pure components.

The algorithm attempts to find the AOD at 0.55 μ m and the surface reflectance at 2.12 μ m that exactly matches the MODIS measured reflectance at 0.47 μ m. This step is done by looping over the mode weight (η) in Eq. 2.8.

3) The final step of this procedure is to find the right fine mode weight (η) from the best solution of the simultaneous relations (2.9). There will be some error, ε , at $0.66 \mu\text{m}$. The best solution is the one where the error at $0.66 \mu\text{m}$ is minimized.

$$\begin{aligned}\rho_{0.47}^m - \rho_{0.47}^* &= 0 \\ \rho_{0.66}^m - \rho_{0.66}^* &= \varepsilon \\ \rho_{2.12}^m - \rho_{2.12}^* &= 0\end{aligned}\tag{2.9}$$

The flowchart with the details of the MODIS Over Land Collection 5 algorithm can be seen in figure 3.1.

In summary, the output products are AOD at $0.55\mu\text{m}$, fine mode weight (η) and the surface reflectance at $2.12\mu\text{m}$ and the residual error ε is also saved for interpretation. However, as we will see in Chapter 3, the operational algorithm also is less than optimum over urban areas. Unlike MISR, the aerosol retrievals are over biased in urban areas and as we will see, this is a direct consequence of the surface albedo estimates MODIS uses in the visible channels.

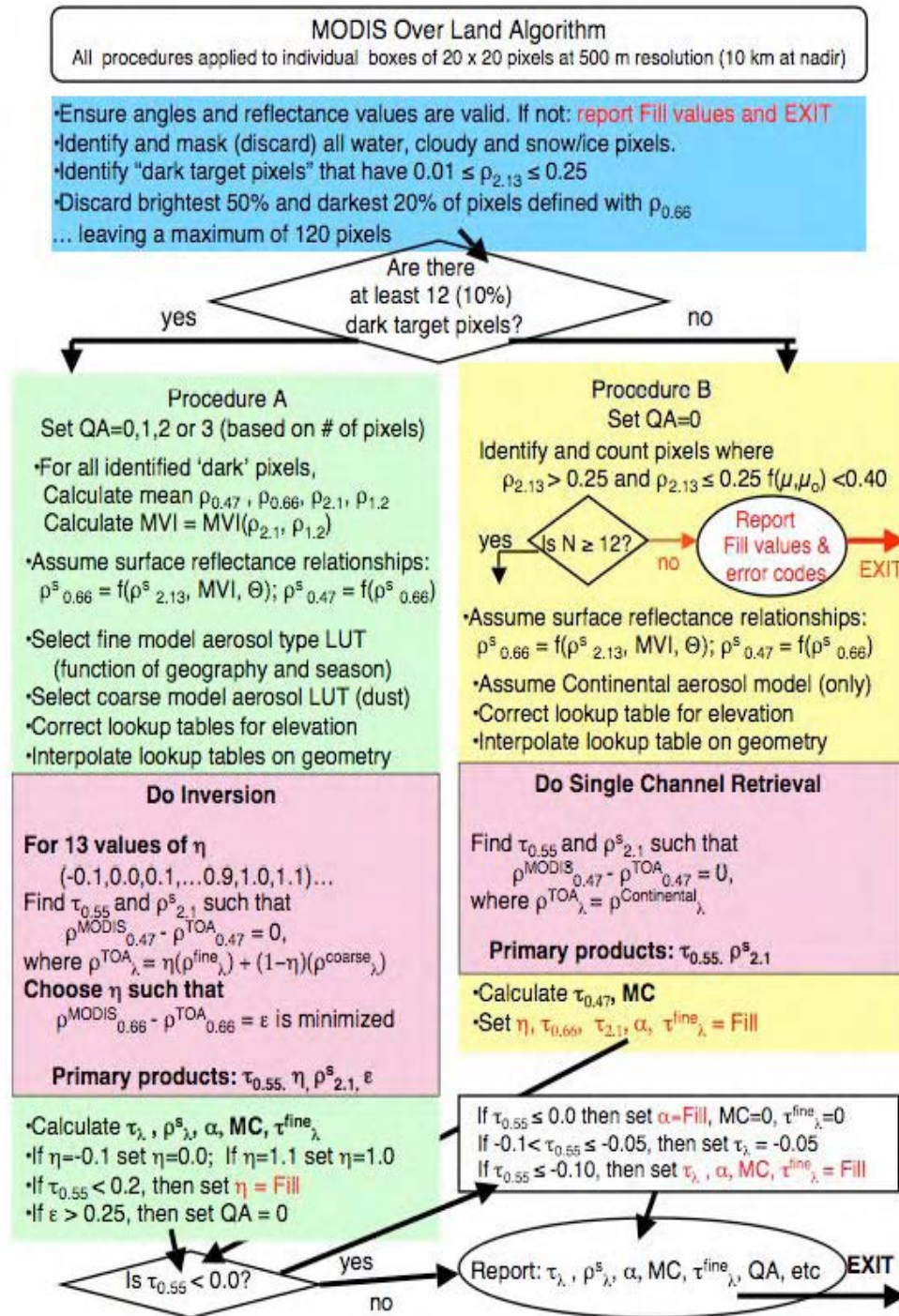


Figure 2.1: Flowchart of the MODIS Over Land collection 5 algorithm.

In order to improve aerosol retrieval over urban areas, surface models which allowed for non vegetative scenes are needed. In particular, there was a significant effort to use AERONET aerosol retrievals (which will be discuss in greater detail in chapter 3) together with MODIS satellite observations to produce more accurate VIS/SWIR relationships that are robust over different surface types. Unfortunately, urban sites make up a very small fraction of the AERONET network so these matchups were not significant to the MODIS team in developing the surface models. However, as we will discuss in Chapter 3, there has been recent interest in modifying the surface model for urban areas using simultaneous measurements from sunphotometers and MODIS TOA reflectances.

For example, in Mexico City, such an effort was performed using a set of distributed hand-held photometers as a means to explore different regions of the metropolitan area. In particular, when the data from the MILAGRO (Megacity Initiative: Local and Global Research Observation) project [Castanho et al 2007] was analyzed for AOD retrieval at $10 \times 10 \text{ km}^2$ spatial resolution , it was found that the VIS-SWIR surface ratio between 660 and 2120 nm value was found to be best described by a numerical value of (0.73). However, when examined at higher resolutions (i.e. $1.5 \times 1.5 \text{ km}^2$), a high variability of the surface reflectance ratio of VIS (660)/SWIR (2120) was found which is significant enough to increase the uncertainties in the AOD values retrieved from satellite at the higher spatial resolution. These values are clearly not accessible to the Collection (5) models which result in a maximum peak of approximately 0.59 for the 660 nm channel.

CHAPTER 3

Developing regional Surface Models

In this chapter, we discuss efforts made by Dr. Min Min Oo to improve satellite retrieval over urban areas. As discussed earlier, in collection 4, a fixed relationship between the visible surface reflectances (470 nm and 660 nm) and the 2120 nm reflectance measured at top of atmosphere was used. These were $\rho_{0.47} / \rho_{2.12} = 0.25$, and $\rho_{0.66} / \rho_{2.12} = 0.5$. In Collection 005 these relationships are $\rho_{660} = f_1(MVI, \Theta) \rho_{2120}$, $\rho_{470} = f_2(MVI, \Theta) \rho_{660}$ where the modified vegetation index $MVI = (\rho_{1240} - \rho_{2120}) / (\rho_{1240} + \rho_{2120})$ is sensitive to vegetation and Θ is the scattering angle. The use of the MVI index instead of the conventional NDVI is due to the relative insensitivity of the MVI to aerosols. These new dependencies permit different relationships depending on the degree of vegetation of the surface. They also account for some of the scatter introduced by BRDF effects.

3.1 Preliminary validation result of MODIS derived AOD with AERONET retrieved AOD

In order to validate MODIS AOD retrieval using inter-comparisons of MODIS and AERONET data, it is very important that only spatially and temporally homogeneous datasets are used. Since the MODIS retrieval is often cloud contaminated, and the surface

reflectance (albedo) itself is spatially variable, the spatial homogeneity of the CIMEL dataset was determined by using the condition that a 5 hour interval surrounding the MODIS inter-comparisons is stable within 10%. The retrievals of AERONET were then compared against two optical depth retrievals from MODIS using Collection (4). The first value is the aerosol optical depth obtained closest to the sky radiometer position which would be sensitive to MODIS algorithm bias due to urban scenes while the second value is the minimum AOD measured by MODIS (in 2001) within an 80km box around the sky radiometer. It is reasonable to expect that the minimum AOD should be far less sensitive to algorithm bias. The results of this inter-comparison are illustrated in Figure 3.1.

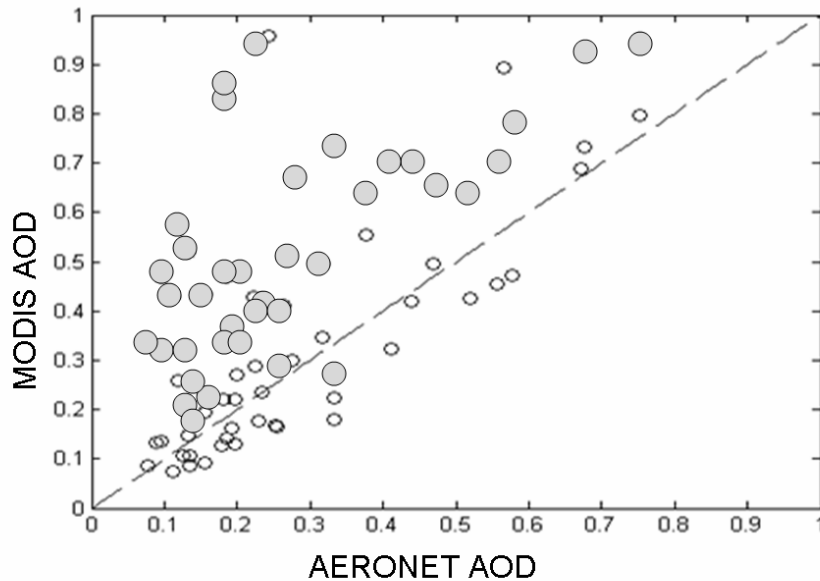


Figure 3.1: Minimum distance MODIS AOD (large opaque circles) and minimum MODIS AOD (small transparent circles) versus AERONET AOD. The large circles correspond to MODIS AOD retrieved over New York City, and the small circles correspond to MODIS AOD retrieved over greener areas to the north the city. The broken line represents the ideal one to one line.

In particular, the MODIS retrievals for the nearest pixel measurements are strongly overestimated. On the other hand, the MODIS minimum AOD measurements are much more accurately correlated to the AERONET measurements. The reason for this is clearly deduced by determining the spatial locations within the 80km box where the minimum AOD is measured.

Figure 3.2 shows the 80 km box enclosing the AERONET site at the City College of New York (marked by the star). The top location marked in figure 3.2 shows the minimum AOD occurred in the highly vegetated and less populated areas to the north of New York City. Therefore we can state that the direct influence of the surface reflectance assumption is clearly correlated to the bias in the retrieval of AOD over urban scenes. Thus, the need of a land surface reflectance model which is not static and connected to the surface type is crucial.

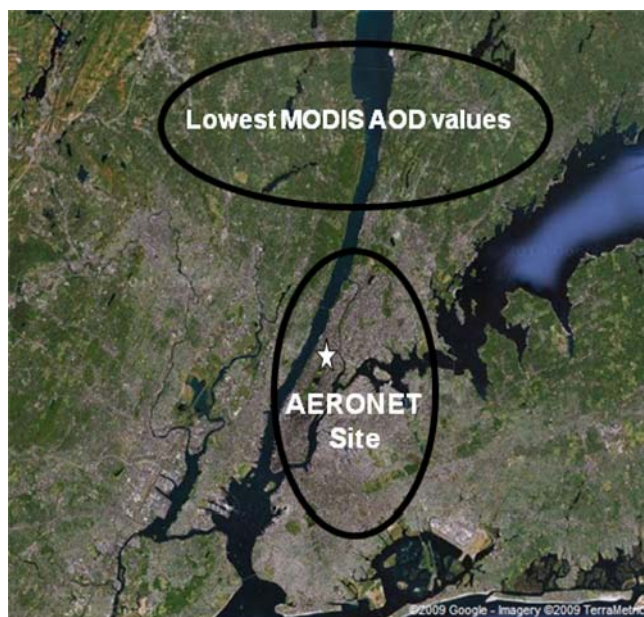


Figure 3.2 Image of New York City and surrounding areas from which the MODIS AOD measurements were taken. The location of the AERONET site is marked by the star. The Minimum values of AOD from MODIS were obtained north of the city on the greener and less densely populated areas to the north of New York City. Image was obtained from Google Maps [www.googlemaps.com].

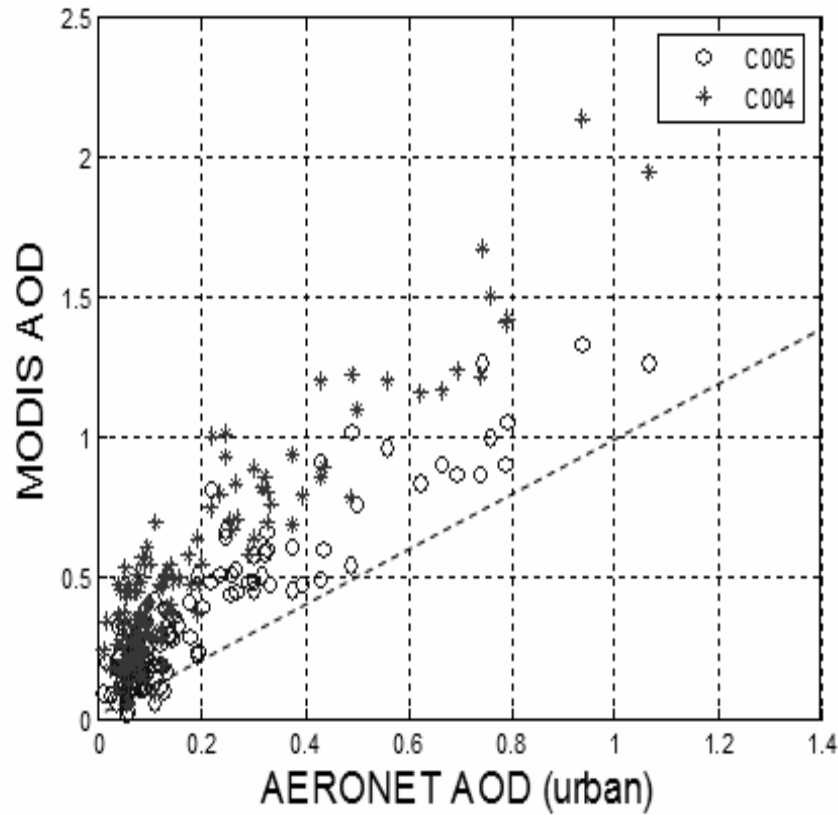


Figure 3.3: Comparison of MODIS retrieved AOD (C004 and C005) at 550nm with AERONET derived AOD (CIMEL sun-photometer data acquired in New York City (the City College of New York) from 2001 to 2007.

The simplest assessment of the different algorithms is done by simple matchup based on the data filters as described in chapter 2. In Fig 3.3, we compare the MODIS retrieved AOD in two algorithms, C004 and C005, against AERONET AOD derived from CCNY station. Clearly, even with C005 algorithm, significant positive bias can still be found. This is because C005 improvements obtained from global modeling is not accurate for regional urban area like New York City. To get some more insight into why the Collection (5) algorithm provides only partial improvement, the urban area reflection

ratio obtained in the C005 algorithm for both vegetative and urban scenes is plotted in Figure 3.4 below.

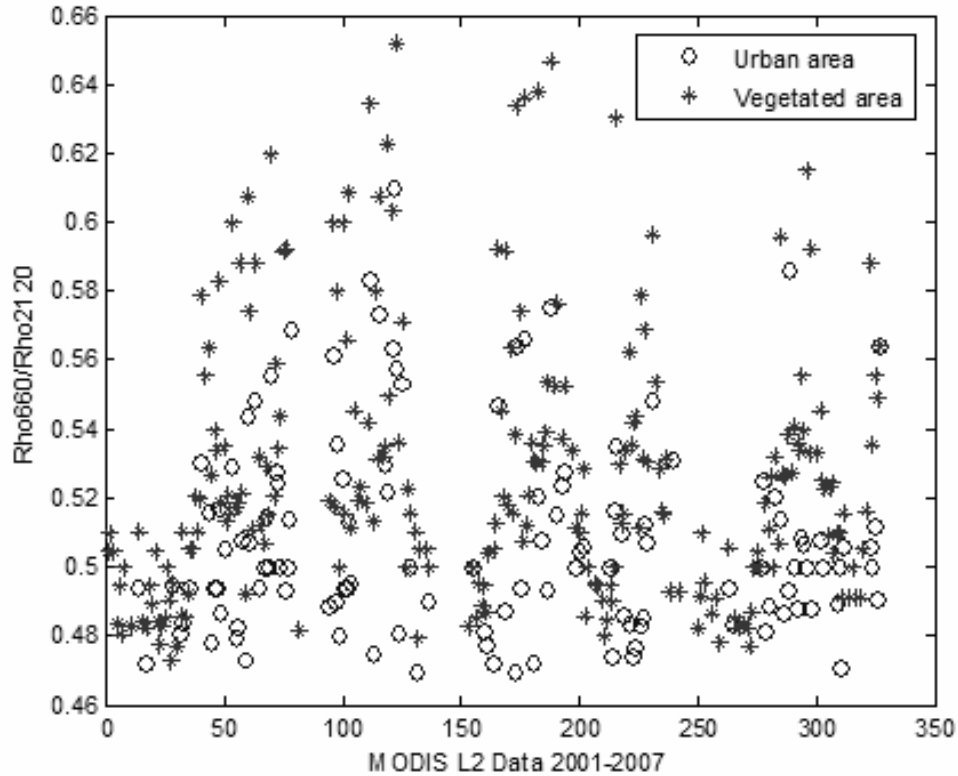


Figure 3.4: MODIS L2 Collection (5) VIS/SWIR surface albedo ratio.

The urban surface reflection ratio of (660nm / 2120nm) derived by MODIS Collection (5) algorithm is approximately 0.55 (downloaded L2 dataset 2001 to 2007). That number is sometimes higher than MODIS Collection (4) algorithm value, which is 0.5. However, as we will see in Chapter 4, it is significantly lower than the number retrieved using high resolution Hyperion Imagery [Gross et al 2005]. In fact, the reason the C005 (un-tuned) algorithm is not that useful for urban retrieval is that the correlations

are only about 10% -20% higher than the static relationship used in the Collection (4) algorithm. The true correlation values are dramatically larger than these VIS/SWIR ratios particularly in heavily urban scenes we will discuss.

3.2 MODIS Surface Albedo Retrieval Procedure: Sunphotometer

In this section, we show how it is possible to generate a per-pixel estimate of the surface VIS/SWIR ratio from high spatial resolution MODIS L1B data MODIS (over the period from 2001-2007) in combination with a-priori knowledge of the aerosol properties obtained from AERONET retrievals [Oo et al., 2010].

Although the MODIS defined LUT has 4 fine aerosol models (Continental, Generic, Smoke, Urban) and a Dust aerosol model, the C005 ATBD (Algorithm Theoretical Basis Document) shows that based on extensive AERONET measurements for the North East region, the fine aerosol model is always defined as belonging to urban aerosol type [Remer et al., 2003; Levy et al, 2007 a]. Therefore, in obtaining the surface model, we will also assume the aerosols are of urban type to interpret our comparisons and not have them depend on aerosol classification errors from C005. The physical and optical properties of the 5 aerosol models are defined in the C005 ATBD [Remer et al., 2003].

During the comparison, we only use atmospheres which we could verify a-priori as fine mode. In fine mode dominated aerosol conditions, the SWIR reflectance at TOA is totally due to the ground since the SWIR wavelength is transparent to the fine mode aerosol atmosphere [Kaufman et al, 1997]. Therefore we can approximate the surface reflectance at $2.12 \mu\text{m}$ as the TOA reflectance and not concern ourselves with aerosol

contributions at the SWIR during surface retrieval. To identify fine mode dominant cases, we select only comparisons when the angstrom coefficient $\alpha > 1$ (as defined in equation (3.1)), and to further reduce errors due to misclassification of aerosols, we limit

$$\text{the AOD to } \tau_{0.55\mu\text{m}} < 0.2. \quad \alpha = \frac{\log(\tau_{675\text{nm}} / \tau_{1020\text{nm}})}{\log(1020 / 675)} \quad (3.1)$$

Moreover, to ensure horizontal homogeneity in the aerosol during the surface retrieval, we require at least 10 cloud cleared points to exist in AERONET measurement in the 4 hour interval (approximately 2 hours before and 2 hours after MODIS/Terra passing time) surrounding the MODIS observations and relative standard deviation in the AOD to be less than 20%. Using a reasonable assumption of mean wind speeds between 10 and 20 km / hr, 4 hour time stability can be converted into 40-80km spatial homogeneity [Anderson et al 2003]. These filters help to reduce possible errors in our assumption that the aerosols are of the urban type and to ensure that regions in the vicinity of the AERONET site are described approximately by the same aerosol model.

To use the AERONET AOD data, we must be able to inject this information directly into the radiative transfer code. Assuming a Lambertian surface albedo, the surface reflectance can be written as

$$\rho_{\lambda}^{surf}(\theta_0, \theta, \Delta\phi, \tau) = \frac{\rho_{\lambda}^{TOA}(\theta_0, \theta, \Delta\phi, \tau) - \rho_{\lambda}^{path}(\theta_0, \theta, \Delta\phi, \tau)}{\bar{s}(\theta_0, \tau)(\rho_{\lambda}^{TOA}(\theta_0, \theta, \Delta\phi, \tau) - \rho_{\lambda}^{path}(\theta_0, \theta, \Delta\phi, \tau)) + T_{\lambda}^{up}(\theta_0, \theta, \tau)T_{\lambda}^{dn}(\theta_0, \tau)} \quad (3.2)$$

where: ρ_{λ}^{TOA} is the measured top of atmosphere reflectance from MODIS L1 B cloud free data (where λ = MODIS band 3 ($0.47\mu\text{m}$) and MODIS band 1 ($0.66\mu\text{m}$)), ρ_{λ}^{path} is the atmospheric path reflectance, $\bar{s}$ is the atmospheric albedo, T_{λ}^{dn} is the downward transmission, and T_{λ}^{up} is the upward transmission, θ_0 , θ , $\Delta\phi$ are the solar zenith angle, sensor (satellite) zenith angle and relative solar sensor azimuth angle respectively [Remer

et al, 2003; Levy et al, 2007 b]. Ingesting the AERONET AOD and MODIS urban aerosol phase function is sufficient to obtain the surface reflections for both visible channels independently. Furthermore, since the aerosols are fine mode dominated, the ground reflectance at $2.12\ \mu\text{m}$ is directly obtained allowing us a measure of the VIS-SWIR surface reflectance ratio.

Using the equation (3.2), the surface reflectances can be calculated. The results of these calculations as a function of the scattering angle are shown in Figures 3.5a, 3.5b, and 3.5c for 3 spatial resolutions (10 km, 3 km, and 1.5 km respectfully) surrounding the City College of New York (AERONET site CCNY). We also masked the inland water area contamination to our derived surface reflectances, since CCNY is so close to Hudson River. More details of the masked can be found at Oo et al, 2010. We first note that the calculated VIS/SWIR reflectance ratios are higher than those implemented from MODIS operational algorithm defined VIS/SWIR reflectance ratios and these ratios are in good agreement with VIS/SWIR ratios result from Hyperion data for New York City urban area [Gross et al 2003]. Furthermore, we do not detect any meaningful trend behavior as a function of angle even though the data is aggregated over the entire multi-year training period.

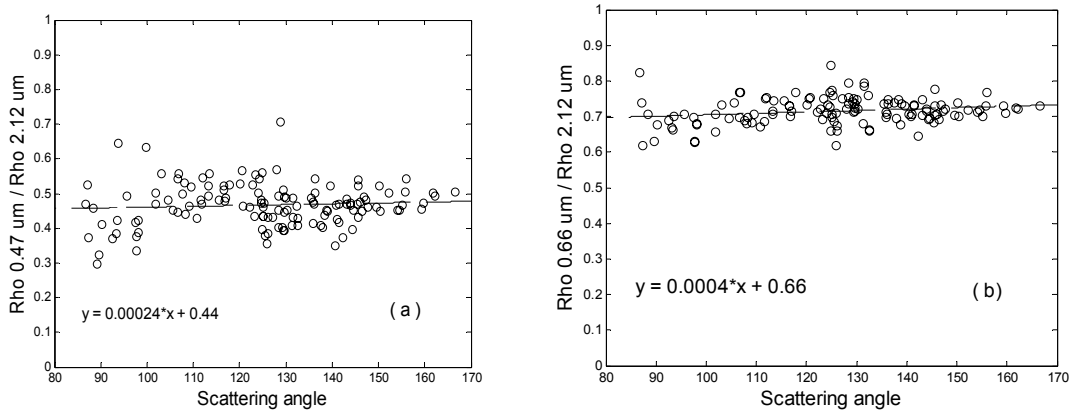


Figure 3.5a Surface reflectance ratio of $0.47 \mu\text{m} / 2.12 \mu\text{m}$ (fig. (a), left) and $0.66 \mu\text{m} / 2.12 \mu\text{m}$ (fig. (b), right), with respect to scattering angle (x axis). These mean reflectance ratios were calculated using $10 \text{ km} \times 10 \text{ km}$ boxes surrounding the nearest pixel to AERONET site at CCNY from 2001 to 2006 with fine mode aerosol dominant atmosphere (with inland water areas and rivers masked).

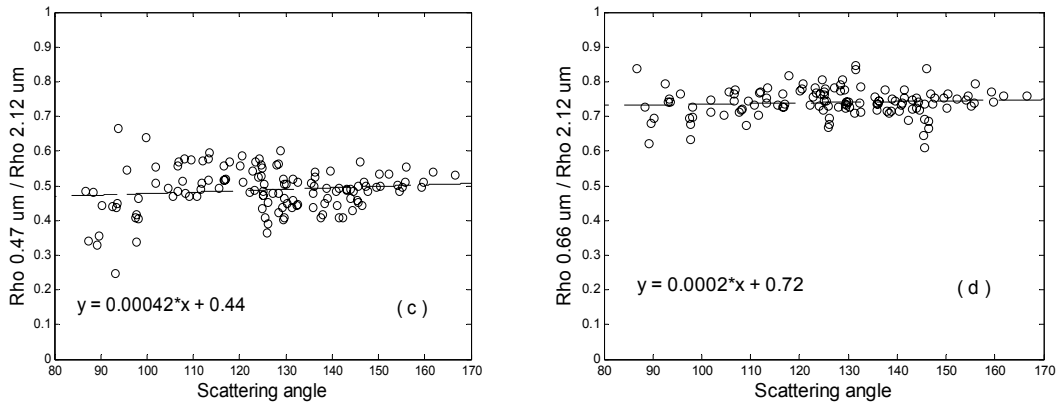


Figure 3.5b Surface reflectance ratio of $0.47 \mu\text{m} / 2.12 \mu\text{m}$ (fig. (c), left) and $0.66 \mu\text{m} / 2.12 \mu\text{m}$ (fig. (d), right), with respect to scattering angle (x axis). These mean reflectance ratios were calculated using $3 \text{ km} \times 3 \text{ km}$ boxes surrounding the nearest pixel to AERONET site at CCNY from 2001 to 2006 with fine mode aerosol dominant atmosphere (with inland water areas and rivers masked).

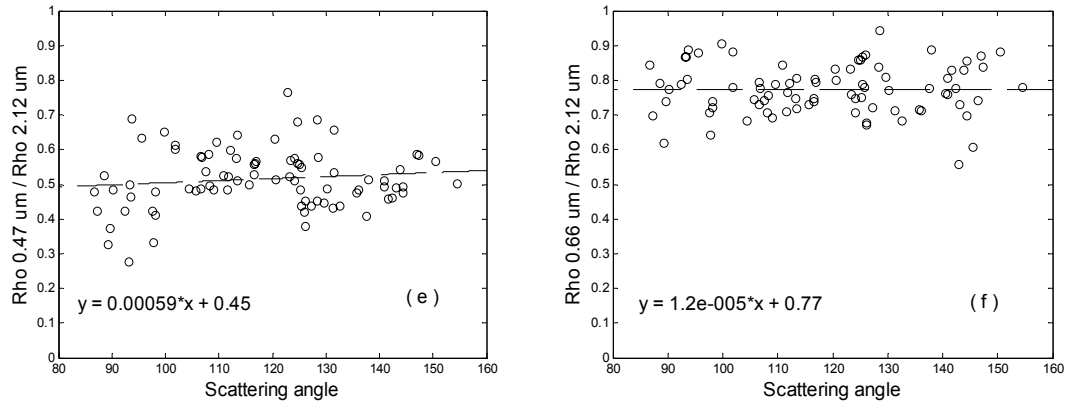


Figure 3.5c. Surface reflectance ratio of $0.47 \mu\text{m} / 2.12 \mu\text{m}$ (fig. (a), left) and $0.66 \mu\text{m} / 2.12 \mu\text{m}$ (fig. (b), right) with respect to scattering angle (x axis). These mean reflectance ratios were calculated using $1.5 \text{ km} \times 1.5 \text{ km}$ boxes surrounding the nearest pixel to AERONET site at CCNY from 2001 to 2006 with fine mode aerosol dominant atmosphere (with inland water areas and rivers masked).

In fact, any trend observed is significantly smaller than the fluctuations around the trend line. It should also be pointed out that the approach was validated against vegetation targets returning values consistent with the C004 and C005 values.

Using the above approach and filtering the data so only homogeneous conditions are available, the AERONET AOD can be assumed to be representative of the region and the surface spectral ratio coefficients can be extended over the region. In particular, we determine the VIS/SWIR surface reflectance ratio as the mean correlation coefficient of each pixel (averaged over all angles). The results of VIS/SWIR surface reflectance ratios at 1.5 km resolution are shown in Figure 3.6 and clearly illustrate the need for a regional surface correlation model.

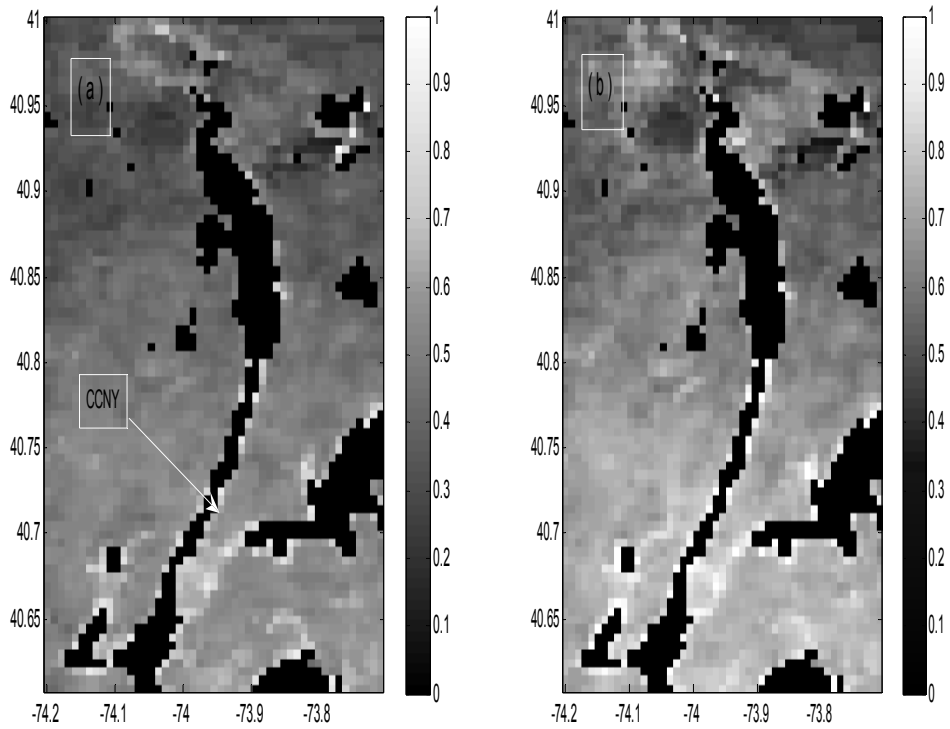


Figure 3.6: VIS/SWIR correlation coefficient ratios of $0.47/2.12\ \mu\text{m}$ ((a) left panel) and $0.66/2.12\ \mu\text{m}$ ((b) right panel) in nearby New York City area.

The regional map extends from 40.61N latitude to 41.4N Latitude, $74.2\ \text{W}$ longitudes to $73.71\ \text{W}$ longitudes. Basically the VIS/SWIR ratios are significantly higher in the urban area compared to the vegetated areas, as shown in figure 3.6. In the northern vegetated areas, the VIS/SWIR $0.47/2.12\ \mu\text{m}$ ratio is ~ 0.25 and the $0.66/2.120\ \mu\text{m}$ ratio is ~ 0.5 to 0.6 which agrees well with the MODIS Collection (5) documentation but the VIS/SWIR ratio in the urban area is much higher. This is particularly clear in urban scenes on both sides of the Hudson River (i.e., New Jersey and Manhattan).

3.3 Path reflectance method

In an alternative approach with high resolution images, a surface AOD retrieval may not be necessary. In particular, we show that by suitable regression over homogeneous surface types undergoing significant shadowing, it is possible to estimate the spectral correlation coefficients.

As described in section 2, the TOA reflectance ρ^{TOA} is given as

$$\rho^{TOA}(\mu_0, \mu, \Delta\phi) = \rho^{Path}(\mu_0, \mu, \Delta\phi) + \frac{T_d(\mu_0)\rho^s T_u(\mu)}{1 - \bar{S}\rho^s} \quad (3.3)$$

where μ the satellite viewing direction, μ_0 is the solar zenith angle, ρ^{Path} is the atmospheric reflectance, ρ^s is the ground reflectance and $T_d T_u$ are the downward and upward total (direct + diffuse) atmospheric transmission. S is the atmospheric albedo. The path reflectance method is quite simple when applied to low atmospheric optical depth conditions. That condition allows the spherical atmospheric albedo S to be approximated as zero in Eq 3.3. Using that assumption, Eq 3.3 is a simple combination of two parts, the atmospheric path reflectance part and the ground reflectance part

$$\rho^{TOA}(\mu_0, \mu) \approx \rho^{Path} + T_{atm}(\mu_0, \mu)\rho^s \quad (3.4)$$

In the SWIR wavelength, the atmospheric path reflectance is approximately equal to zero and the upward transmittance and downward transmittance approximately equal to one for low atmospheric optical thickness and/or fine mode aerosol dominates in the atmosphere. Therefore in this condition, the TOA reflectance signal is directly coming from surface reflectance in SWIR wavelength while the atmosphere is almost transparent.

$$\rho_{SWIR}^{TOA}(\mu_0, \mu) = \rho_{SWIR}^{Path}(\mu_0, \mu) + \frac{T_d(\mu_0)\rho_{SWIR}^s T_u(\mu)}{1 - \bar{S}\rho_{SWIR}^s} \approx \rho_{SWIR}^s \quad (3.5)$$

Equation 3.5 shows that TOA reflectance in SWIR is approximately equal to the surface reflectance. In addition, assuming a linear correlation exists between the VIS and SWIR channels surface reflectance, according to Kaufman [4], Eq. 3.3 can be written as

$$\rho_{vis}^s = k\rho_{SWIR}^s = k\rho_{SWIR}^{TOA} \quad (3.6)$$

Hence under low optical thickness atmospheric condition, the TOA reflectance in the VIS wavelength is simply:

$$\rho_{vis}^{TOA} \approx k\rho_{SWIR}^{TOA} + \rho_{vis}^{Path} \quad (3.7)$$

When SWIR reflectance closes to the limit of zero, the TOA reflectance in the VIS wavelength is the only signal contributing to the atmospheric path reflectance.

$$\rho_{SWIR}^{TOA} \rightarrow 0 \Rightarrow \rho_{vis}^{TOA} = \rho_{vis}^{Path} \quad (3.8)$$

Therefore, the regression plot between the TOA reflectance in the VIS and SWIR will be linear. In particular, the slope of the regression line will include the correlation coefficient of VIS-SWIR surface reflectance through $k=cT_d(\mu_0)T_u(\mu)$. On the other hand, the correlations can have non-zero offsets which make this approach less suitable. However, using this technique is simple to employ as a first estimate to observe the differences in correlation coefficients of VIS/SWIR surface reflectance of different surface type (urban and vegetation areas).

3.4 Path reflectance method using time series data

Another related approach which can be used to estimate the Spectral reflection correlation coefficients is to use time series data. This approach has the benefit that it can be applied to lower resolution sensors such as MODIS. This approach is similar to that

employed by the Geostationary Aerosol and Smoke Product (GASP) [Level-2 MODIS; Holben et al 2001; AERONET network] that has now been made operational and provides data through the NOAA – CLASS Data system.

For our purpose, the most critical aspect is the development of the GASP algorithm is the generation of realistic surface estimates based on multiple time observations of the same pixel and monitoring this pixel over sufficient times to see cases where the atmosphere loading is sufficiently small. In the GASP algorithm, this is done by first creating a composite clear-sky background image for each time of day which is equivalent to a fixed solar zenith angle for a given pixel. This composite image is generated by collecting images over the previous 28 days including the current image, and the selecting the second darkest pixel at each location over that time period. The choice of using the second darkest pixel instead of the darkest is to avoid or reduce cases where cloud shadow contamination might arise [Knapp et al, 2005]. Although the choice of 28 days is fixed statically operationally, it should be pointed out that such rules may not be optimal and can vary seasonally and geographically seasonally. In particular, the choice is made as a way of balancing between a longer time period which increases the chances of finding an image with clear sky conditions and a shorter time period to increase the probability that the underlying surface reflectivity is stable. For example, as seen in processing the MODIS surface BRDF product for both urban and non urban scenes, we see a significant improvement in stability of the spectral ratios when compared to vegetation. This stability allows us to extend the time period for observations while assuming that the spectral ratios are stable.

One approach that can be used is to use the fact that if we plot the TOA VIS versus the TOA SWIR reflectances and then overlay them with the Rayleigh atmosphere.

$$\rho_{TOA}^{Rayleigh} \approx \rho_a^{Rayleigh} + T_d^{Rayleigh} T_u^{Rayleigh} c_{2130}^{vis} \rho_{TOA}^{2130} \quad (3.9)$$

In particular, the Rayleigh plots are calculated for the appropriate geometric variables for sun and satellite with different spectral ratio slopes. The point is that the Rayleigh atmosphere acts as a lower limit to the reflectance values that we can expect for the atmosphere. In this way, by seeing which spectral ratio makes the Rayleigh retrieval a lower bound to the TOA data is the best estimate of the spectral ratio. Unfortunately, sparse data and significant aerosol loading can make it quite difficult to get really accurate measurements. In our case, we used Dec 2008 data for 30 days combining aqua and terra to get 21 observations. Using December as a test has advantages and disadvantages. Clear sky cases are more likely but at the same time, the solar angles are more severe with solar zenith angles ~ 60 degrees. Still, we explored the possibility below.

In looking at the results in figures 3.7a and 3.7b, we use both the global and zoomed view. The black traces identify the Rayleigh TOA calculated for the appropriate angles as identified in eqn. 3.9 with the values of c_{2130}^{vis} going from 0 to 1 in .05 increments. In identifying the correct value, we look for those observations which intersect the Rayleigh atmosphere retrievals and assume the atmosphere for these cases is well approximated by Rayleigh since additional aerosols will cause the TOA reflectance to increase above the Rayleigh range (as most of the pixels clearly do).

We see that the best value is approximately $c_{2130}^{660} \sim .7$ which again is quite good in comparison to our other approaches. At 470nm, the results are a little less impressive

with $c_{2130}^{470} \sim .55$ which is somewhat larger than the value of 0.38 that we seemed to get from the other methods. However, this can be explained by noting that even small contributions of aerosol are enhanced at shorter wavelengths. Adding a small aerosol component much like GASP should provide even better results.

In fact, when the calculations are redone assuming a minimum AOD with an urban atmosphere of 0.02 which is quite reasonable as a minimum AOD for Dec, we obtain better results with $c_{2130}^{660} \sim .65$ and $c_{2130}^{470} \sim .45$ as illustrated in figure 3.8a and 3.8b.

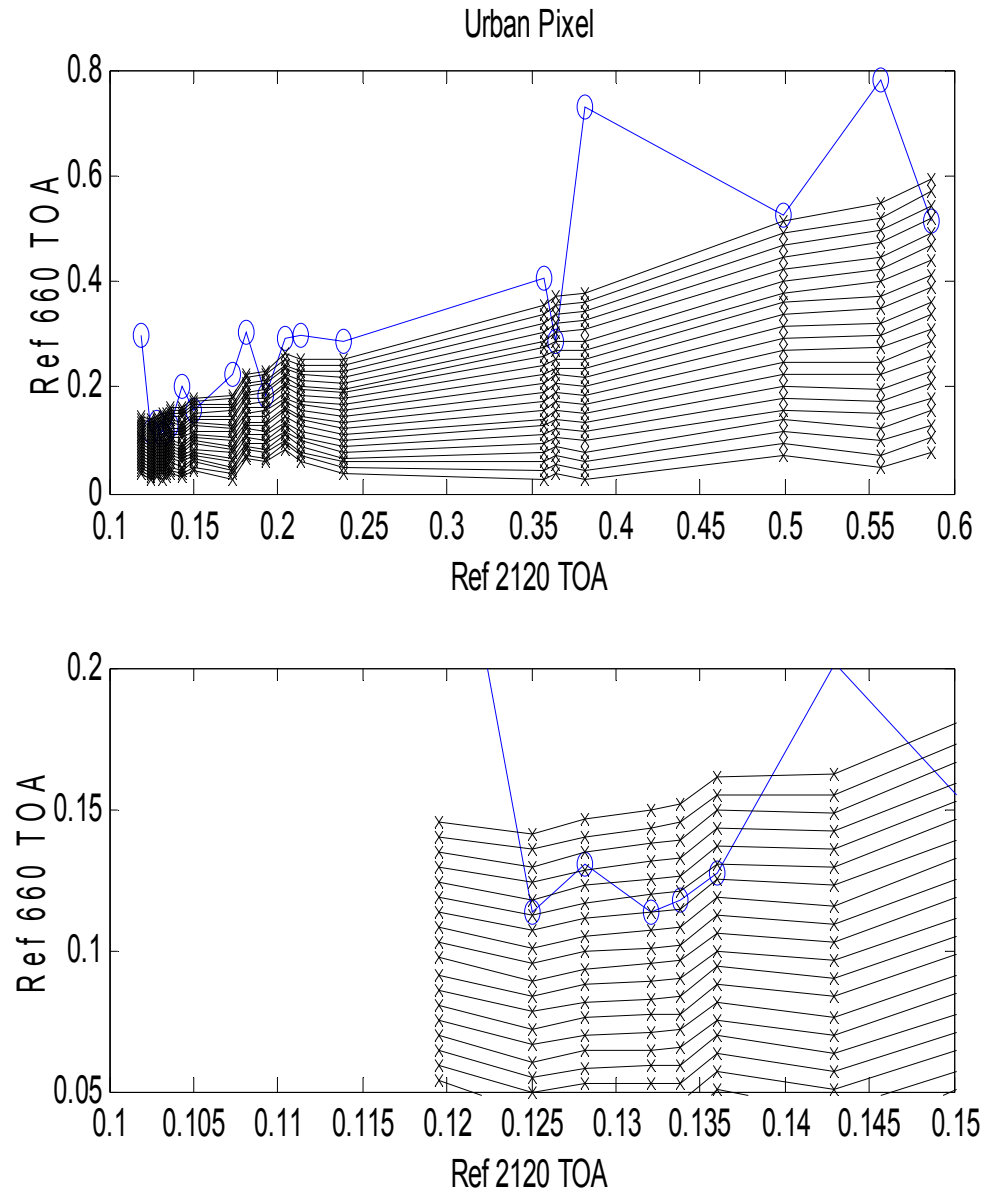


Figure 3.7a TOA reflection at 660 nm over urban pixel overlaid with Rayleigh estimates for different surface correlations. The circles represent the measured values of MODIS over a 30 days period. The lines with x are the calculated reflectances values. Each line was obtained by varying the VIS-SWIR coefficient from 0 to 1.0 in 0.05 increments. (Top) The entire series of measured and calculated TOA reflectances is shown. (Bottom) Zoomed view that shows only the area where the measured TOA reflectance fall within the calculated TOA reflectance.

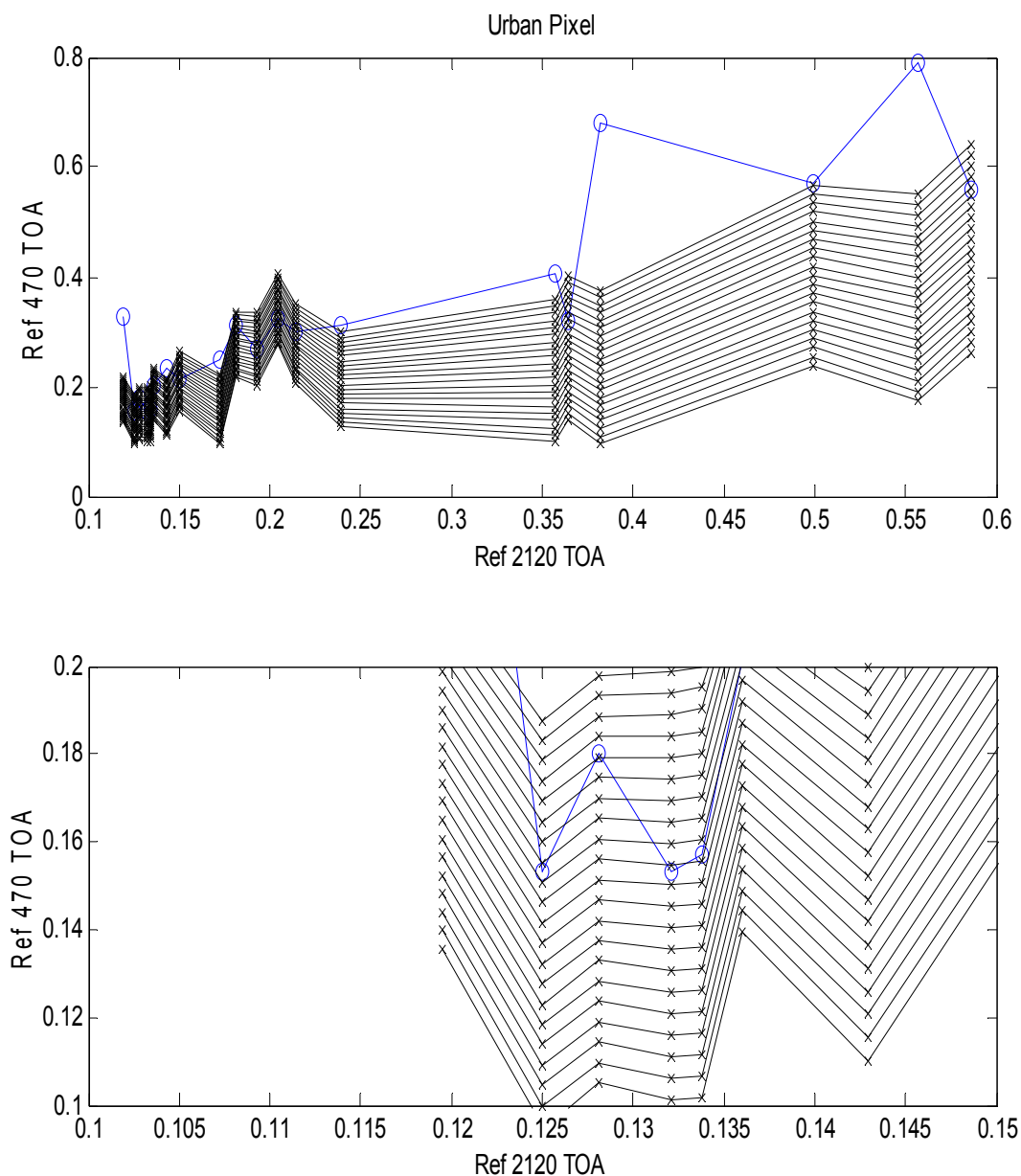


Figure 3.7b TOA reflection at 470nm over urban pixel overlaid with Rayleigh estimates for different surface correlations. The circles represent the measured values of MODIS over a 30 days period. The lines with x are the calculated reflectances values. Each line was obtained by varying the VIS-SWIR coefficient from 0 to 1.0 in 0.05 increments. (Top) The entire series of measured and calculated TOA reflectances is shown. (Bottom) Zoomed view that shows only the area where the measured TOA reflectance fall within the calculated TOA reflectance.

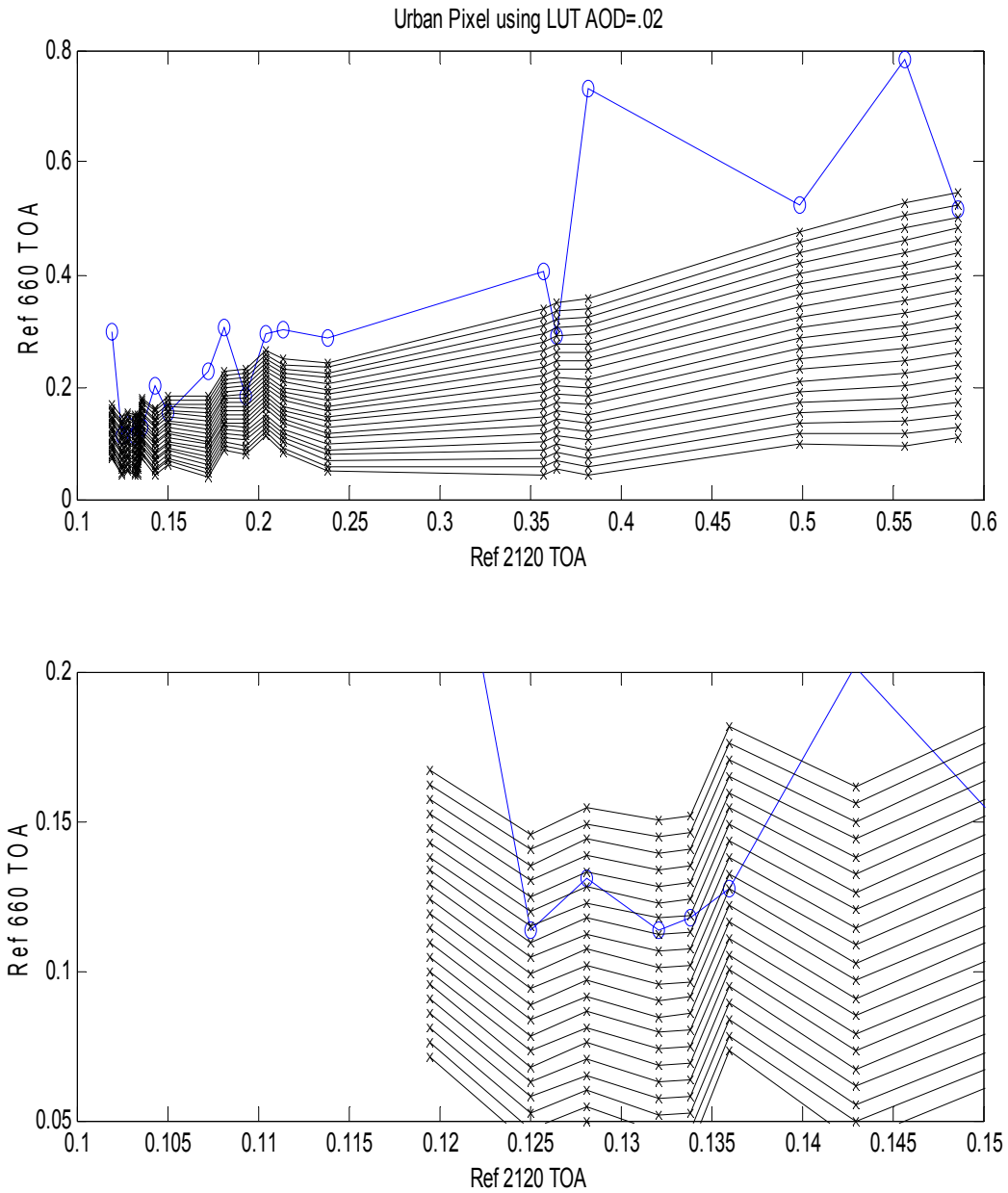


Figure 3.8a TOA reflection at 660 nm over urban pixel overlaid with Rayleigh estimates and small urban aerosol layer with AOD=.02 for different surface correlations. The circles represent the measured values of MODIS over a 30 days period. The lines with x are the calculated reflectances values. Each line was obtained by varying the VIS-SWIR coefficient from 0 to 1.0 in 0.05 increments. (Top) The entire series of measured and calculated TOA reflectances is shown. (Bottom) Zoomed view that shows only the area where the measured TOA reflectance fall within the calculated TOA reflectance.

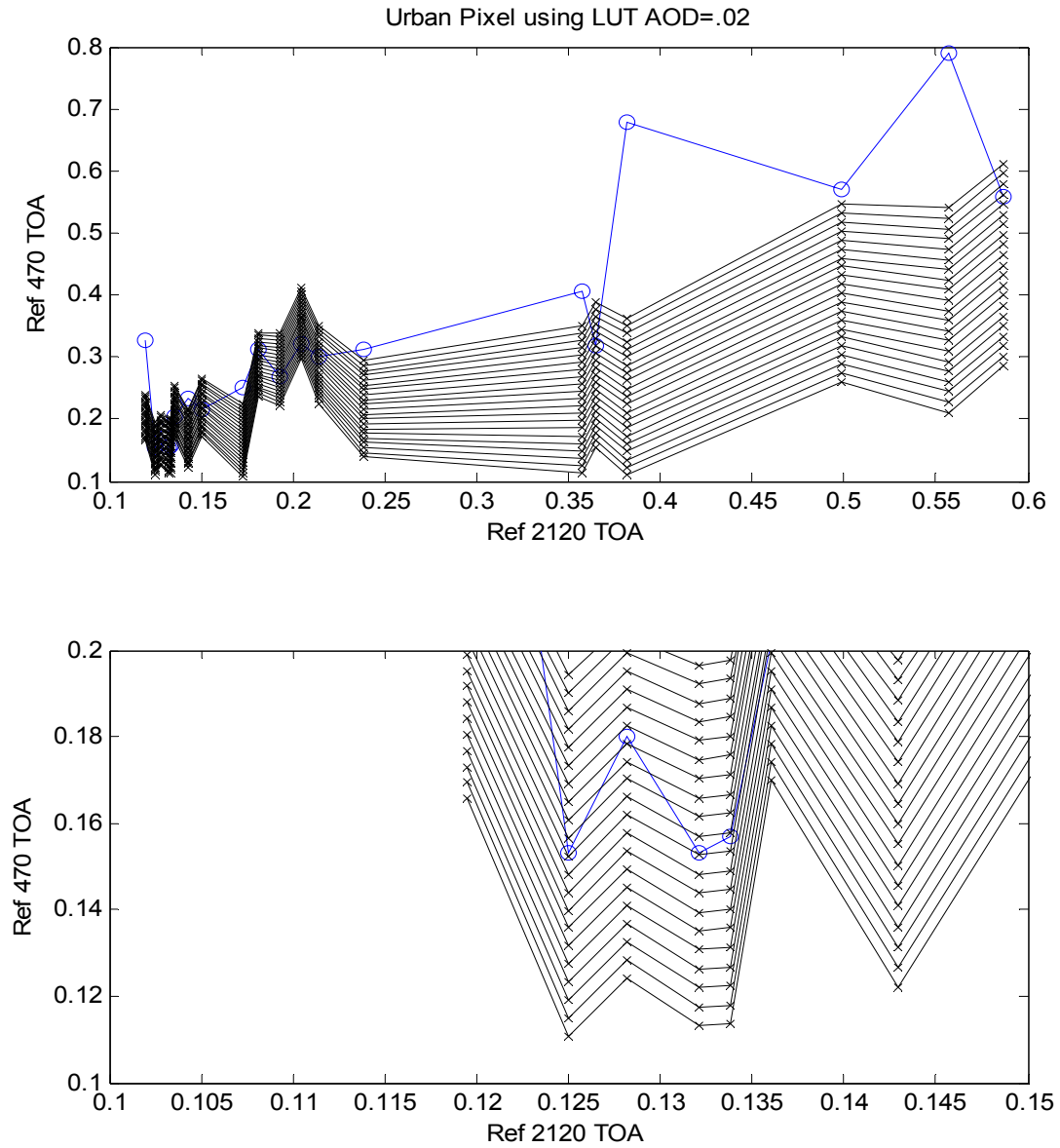


Figure 3.8b TOA reflection at 470nm over urban pixel overlaid with Rayleigh estimates and small urban aerosol layer with AOD=.02 for different surface correlations. The circles represent the measured values of MODIS over a 30 days period. The lines with x are the calculated reflectances values. Each line was obtained by varying the VIS-SWIR coefficient from 0 to 1.0 in 0.05 increments. (Top) The entire series of measured and calculated TOA reflectances is shown. (Bottom) Zoomed view that shows only the area where the measured TOA reflectance fall within the calculated TOA reflectance.

3.5 Preliminary Analysis using the Hyperion Hyperspectral Imager

The above results show in general that the urban area is different than the greener areas north of New York City. However, the direct link between vegetation and urban surfaces and their spectral ratios requires high spatial resolution imagery (i.e Hyperion) that can separate different surface classes. The in-depth details of the processing of the Hyperion sensor and the image used for this analysis will be shown in the next chapter. This 1024 x 256 image over New York City was particularly useful for surface modeling due to the very low AOD observed for this day ($\tau_{500} \leq .05$). In this Hyperion Image, we can visually focus on several distinct regions representing light and heavy urbanization, water, and vegetation (like in Central Park), as we can see in figure 3.9. The segmentation method was later refined by doing supervised classification, as explained in the next chapter.

To illustrate the retrieved correlation coefficients of this method over different scenes, we plot histograms of the correlation coefficients for the three different surface types that can be identified, which are vegetation, rural (light urban) and urban, as seen in figure 3.10. Note that in Central Park (proxy for vegetation) and in light urban areas, the correlation for the 660-2160nm bands is unbiased around the vegetation value of 0.55 although a surprisingly high variability is seen in the results. However, when considering light or heavy urban scenes, the value for the correlation coefficient is significantly higher with values peaking to 0.72 for heavy urban scenes in good agreement with the MILARGO results [Levy, Remer et al 2004].

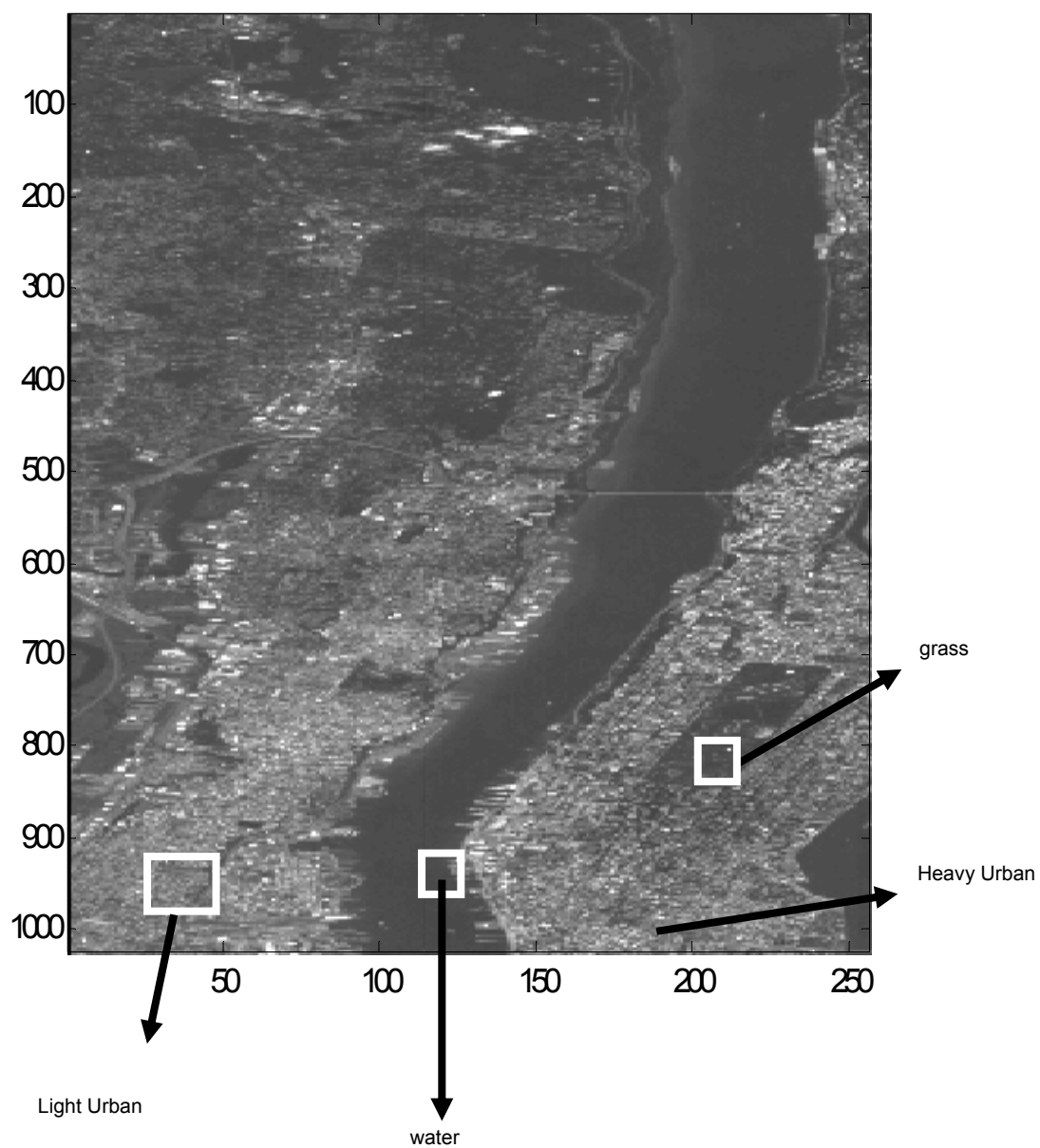


Figure 3.9: Hyperion image of New York City (from 660nm band). Regions of interest include vegetation, Light urban areas, Heavy Urban Area and water.

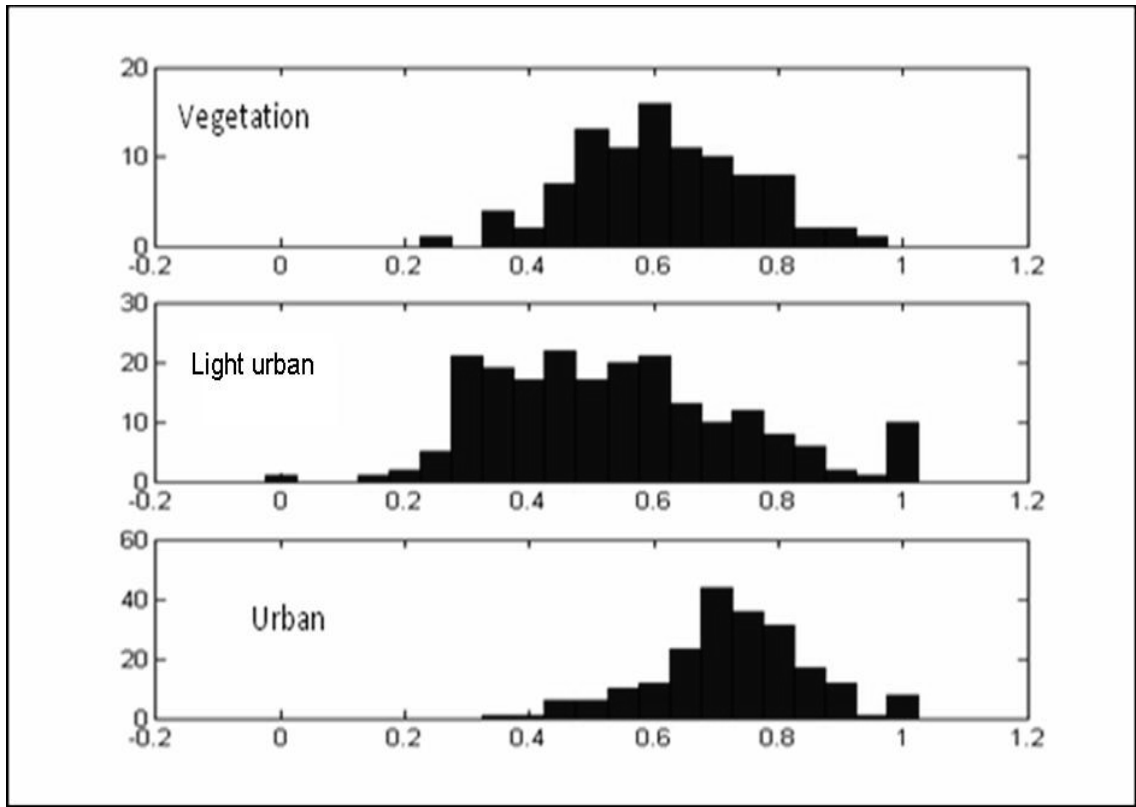


Figure 3.10: Histograms for (Top) vegetation, (Middle) rural and (Bottom) urban pixels. The x-axis contains the values of VIS-SWIR slopes and the y-axis are the number of occurrences.

The results are based on an assumption that the correlation is constant except for intensity fluctuations. Another approach is to probe the surface reflectance directly using atmospheric correction schemes on a pixel-pixel basis using a segmentation method described in the next chapter. After analyzing the image using the results from the segmentation approach, we noted that while the vegetation pixels were quite well described by the MODIS values in both C004 [Level-2 MODIS] and C005 [Holben et al 2001] algorithm, the urban pixels resulted in a significantly higher ratio which is

consistent with the statistical based approach previously used.

3.6 Assessment of Regional Training

In using these urban ratios, derived above by different means, a significant improvement is found as shown in Fig. 3.11. In fact, the uncertainty even at higher resolution is quite small. In all these plots, the operational error bounds of $(\pm .05 \pm .15\tau)$ for AOD retrieval over land were added and show that except for very low AOD values where even small errors in reflection compensation can lead to significant AOD errors, the AOD retrievals are well within these uncertainty bounds.

It was apparent that to find the correlations over different surfaces, we needed a high spatial resolution sensor from which we could isolate the different surfaces. The Hyperion and Landsat 7 (ETM+) sensors provide images with 30 meters pixels resolution. This high resolution allows the visualization and further separation of different surfaces, like vegetation and urban. The following two chapters show the analysis done using imagery from these sensors.

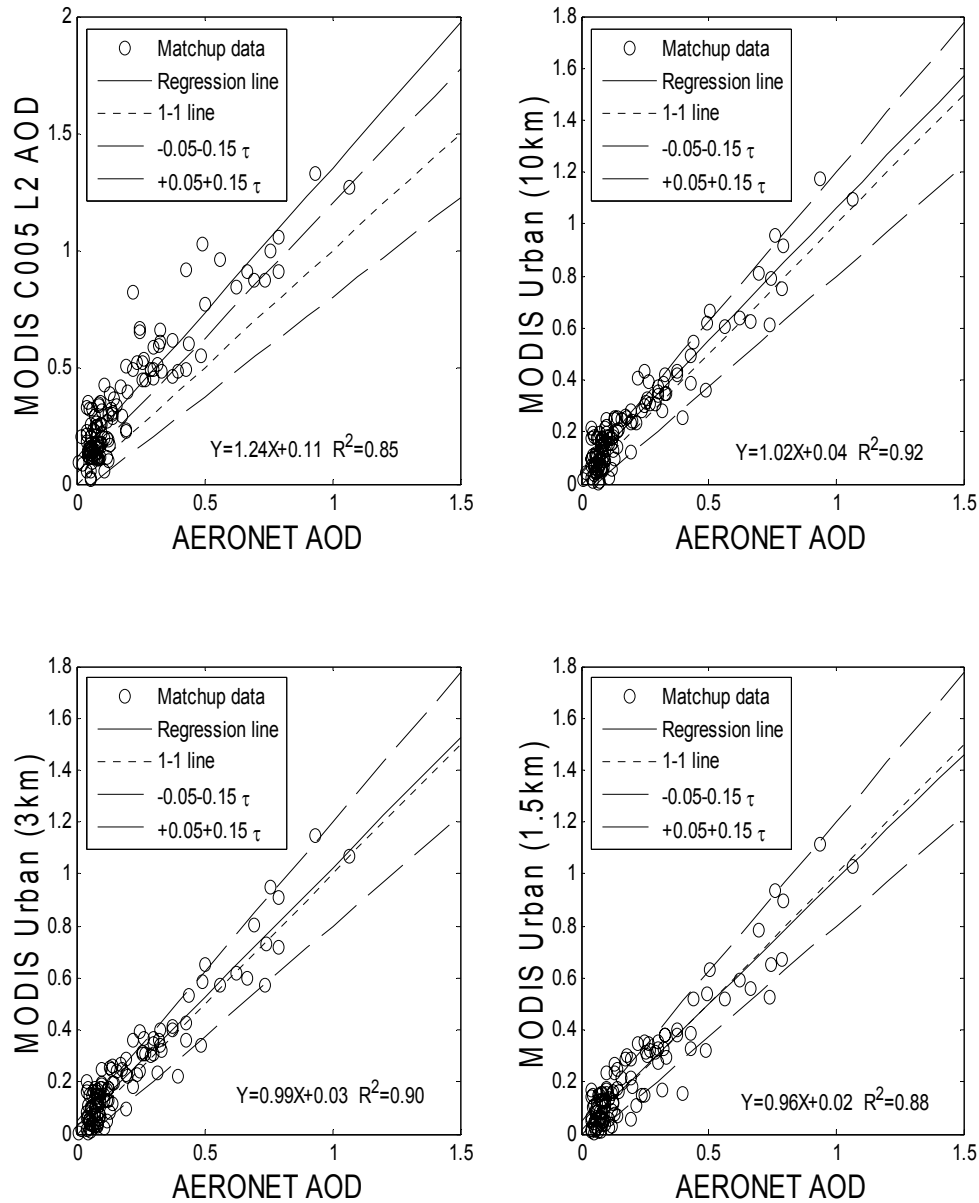


Figure 3.11: Panel (a) MODIS L2 Aerosol Optical Depth at $0.55 \mu\text{m}$ compared with 4 hour ($\sim 2\text{hr}$ before and 2hr after MODIS (Terra) satellite overpass time) average of AERONET aerosol optical thickness. Panels (b), (c) and (d) (using 10 km, 3 km and 1.5 km resolution respectively) are retrieved AOD with new surface reflectance VIS/SWIR ratio plotted versus the average of AERONET AOD.

CHAPTER 4

Hyperion Imagery Analysis

The Hyperspectral Hyperion imager is one of the components of the NASA's EO-1 system, which is part of NASA's Earth Observing System (EOS). It orbits the earth at 750 km following the same track as the Landsat-7 satellite, just 1 minute behind. It generates image swaths of 7.7 km width, with a spatial resolution of 30 meters. The high spatial resolution of the Hyperion image, allows to visually focusing on several distinct regions representing urbanization, and vegetation.

Figure 4.1 shows the sizes of the swaths from the Hyperion and Landsat sensors. Hyperion acquires data in 242 bands ranging from 355 nm to 2577 nm with a 10 nm spectral resolution. Ninety five (95) of these bands could not be used because they either lack data or had a very high amount of artifacts (most of these bad bands were not operating due to a lack of calibration).

Hyperion imagery is now available for free, but when this study started a Hyperion image from the archives cost \$300 dollars, and \$500 for special requests, thus our analysis of Hyperion imagery was limited to a single image. The Hyperion image used in this thesis was a 1024 x 256 image over New York City, which can be seen in figure 4.2. This image did not include the World trade Center (WTC) located further south. It was taken a day after the destruction of the Twin Towers and nearby structures in the WTC. This created an enormous dust cloud and heavy smoke plumes due to the

continuous burning in the rubbles of the massive structures. Even so, the prevailing Northern winds pushed these plumes south and the AOD loading was surprisingly very low in the areas north of the WTC.

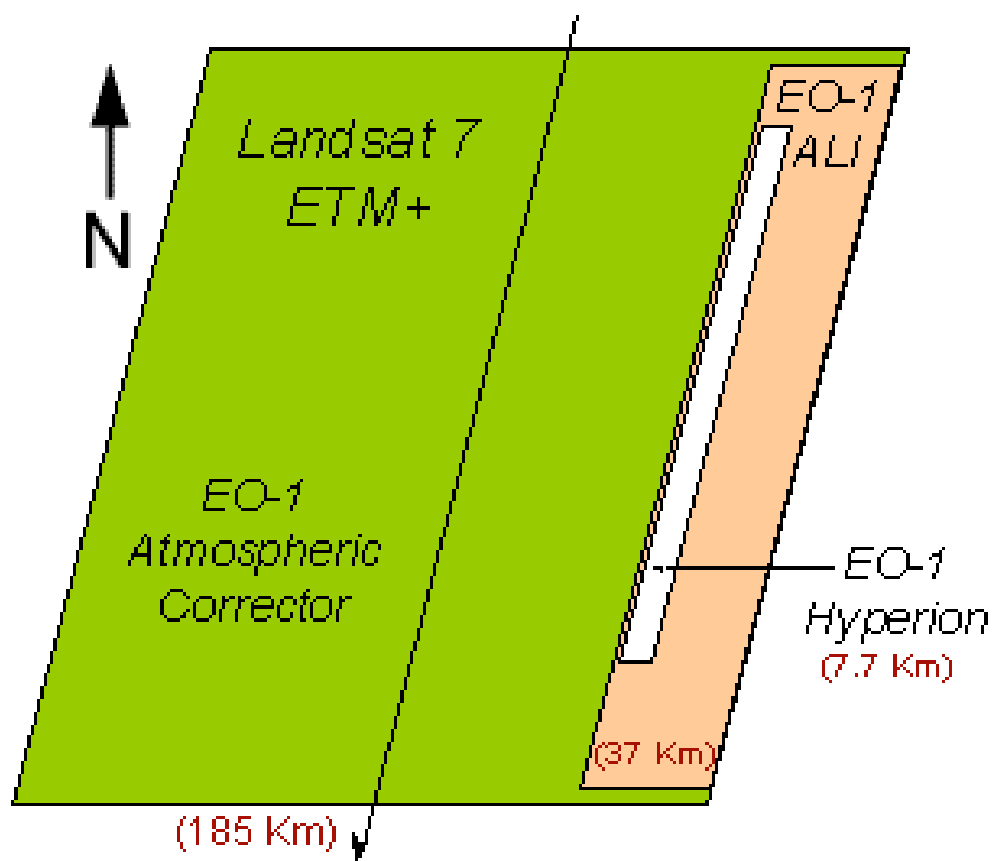


Figure 4.1 Swath width of the images obtained from Hyperion (7.7 km) and Landsat (185 km) sensors.

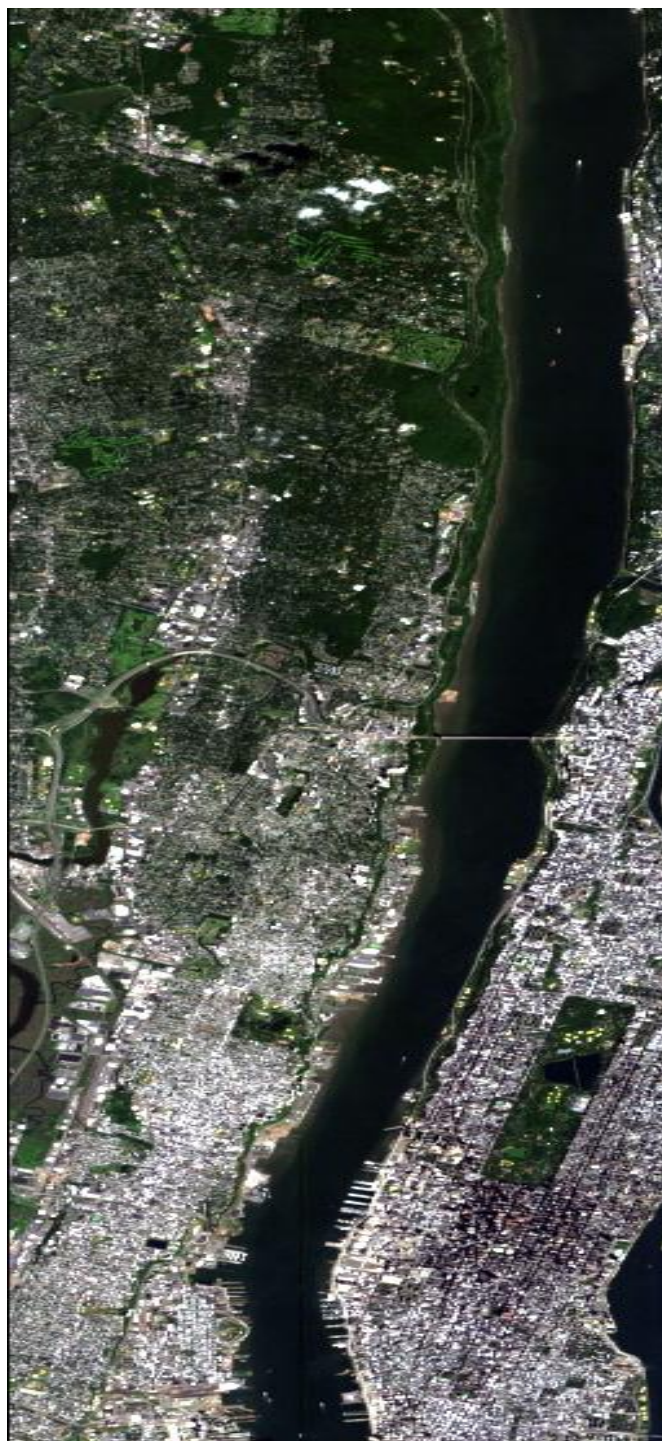


Figure 4.2 Hyperion image over New York City taken on September 12, 2001.

This image was particularly useful for surface modeling due to the very low AOD observed for this day ($\tau_{500} \leq .05$) as shown in the Aeronet data in figure 4.2. The CIMEL instrument from which these AOD measurements were taken was located approximately 6 miles north of the WTC.

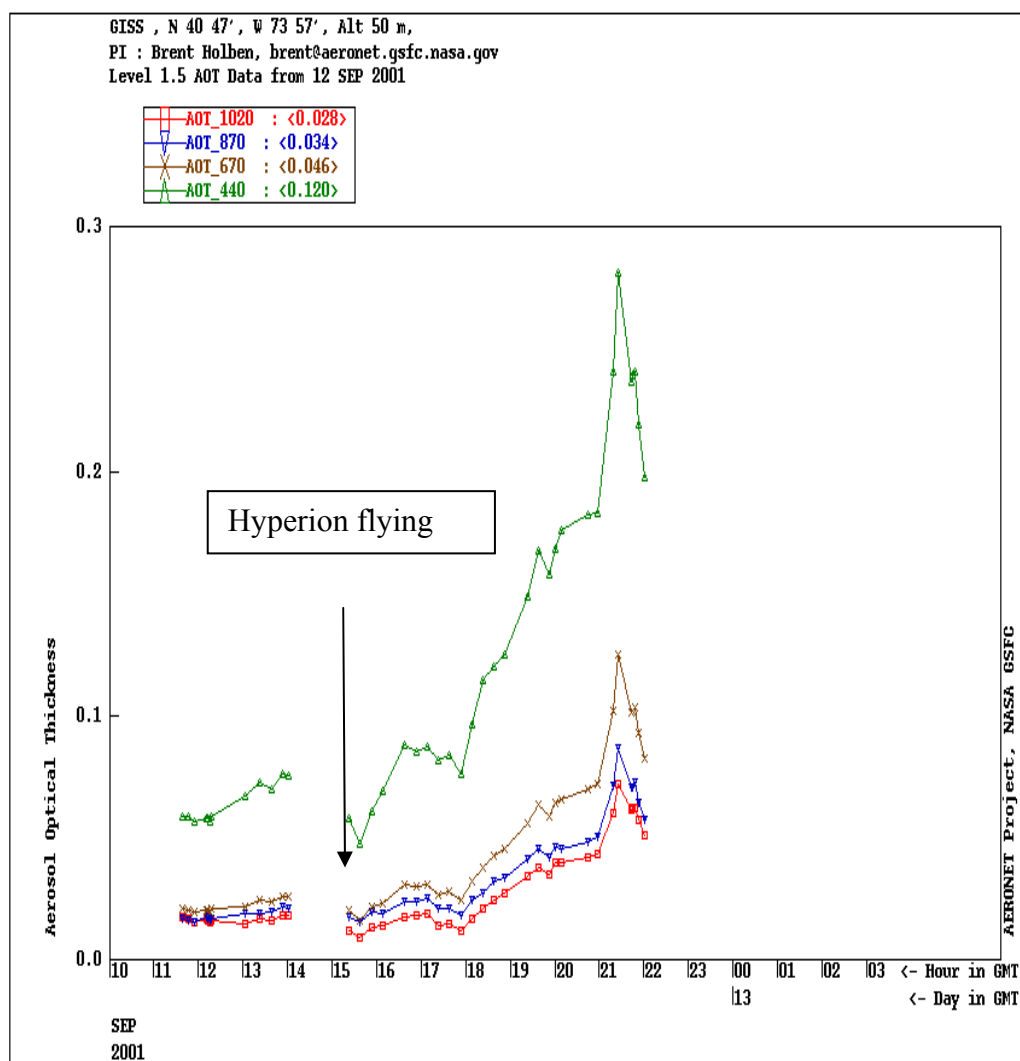


Figure 4.3 Aerosol Optical Thickness over New York City on September 12, 2001. The arrow points to the AOD at the time that the Hyperion image was acquired.

4.1 Hyperion Data Processing and Preparation

The processing of the image was done using ENVI 4.1 and its FLAASH atmospheric correction toolbox. Even though ENVI has some statistical functions, our data analyses were done using the MATLAB 7.1 software. Using the ENVI software, the image files were extracted and saved as ASCII files. The ENVI headings were subsequently removed from each file and then saved as data files, which only contained data values. These data files were loaded into MATLAB as data matrices where they were processed and analyzed. All these software ran on a Microsoft Windows platform.

Even though we had a single image to analyze, we were expecting to acquire a few more. Therefore in order to maintain consistency in the atmospheric correction processing using FLAASH, we tried to find an ideal set of parameters for Hyperion images, which would produce the most accurate and consistent results. At the beginning of this research, we focused extensively into processing the data using different FLAASH parameters. The FLAASH software has a great deal of options to select from as well as many user's input parameters. Many of these values are automatically filled in after entering the sensor type and the image information (i.e. acquisition date and time, latitude and longitude). Other values are contained in the header file of the image data, and are automatically extracted by the software. Since it was not feasible to go through all these parameters, we selected the US Standard atmospheric model and varied the following parameters (the default values were used for the other parameters):

- 1) Initial visibility in km: We used from 20 to 120 km in increments of 10 km.
- 2) Aerosol retrieval: None, or 2-Band (K-T)
- 3) The aerosol model: Rural, Urban, and Tropospheric.

The visibility (V) is inversely related to the AOD and it can be approximated as:

$$V = 3.912 / \beta \quad \text{where } \beta = \text{AOD} / \text{PBL where:}$$

V: Horizontal visibility in kilometers.

β : Aerosol extinction coefficient which is defined as the horizontal optical depth per km.

PBL: Planetary boundary layer, which is the layer of aerosols at the lower atmosphere and may vary from 100 to 3000 meters. Its value is typically about 2 km so that $V = 7.824 / \text{AOD}$.

Therefore, by entering a visibility value, one is providing the aerosol loading value to be assumed for the atmospheric correction. We used visibilities values from 20 km to 120 km in increments of 10 km.

The 2-Band (K-T) aerosol retrieval method is based on the method developed by Kauffman-Tanre that uses dark pixels to determine the visibility to be used for the atmospheric correction. When this option is selected, the FLAASH software disregards the initial visibility input unless it does not find any suitable dark pixels. Our tests did show that regardless of the initial value of visibility entered, the visibility used for the atmospheric correction was always the same. Therefore to continue testing the software using different values of visibilities, we did not use the 2-Band (K-T) aerosol retrieval method.

We then tested the changes in the atmospheric model (urban, rural and tropospheric) for every visibility value previously used. Our results showed that for this image there was little difference between the three models. To determine the optimum atmospheric model for this image, we compared the amount of negative pixels for each set of parameters combination (negative reflectance values could be an indicator of

overcorrection). Most of these early test results were lost in an old hard drive and could not be shown.

We determined that the best options were the 'Urban' model without the 2-Band (K-T) aerosol retrieval. Therefore our analysis of Hyperion data was done with atmospherically corrected reflectance files processed with the FLAASH software using the US Standard atmospheric model, Urban aerosol model, and visibilities ranging from 20 km to 120 km in increments of 10 km.

Our next task was to find a suitable segmentation method to extract the vegetation and urban regions. The following section describes the segmentation procedure.

4.2 Image Classification

Classification methods are techniques used to classify the pixels of an image into clusters of data that have similar statistical values, and thus considered to belong to the same group or class. Classification methods can be accomplished in a supervised or unsupervised manner. In unsupervised classification, the pixels of the images are classified without using any previous knowledge of the number of classes present in the image. The user enters any arbitrary number of classes in which the algorithm will classify the data. ENVI has two unsupervised classification methods, the IsoData and K-Means, neither one of them gave a good classification.

In supervised classification the user visually inspect the image and select samples (called regions of interest) from the features to be classified. The supervised algorithms classify the image into classes that share similar statistical values to the regions of

interest. ENVI has several supervised classification tools. We tried all of them, and the best results were obtained from the parallelepiped method.

The parallelepiped is the simplest and fastest of the supervised classification methods. Each class can be represented as a normal distribution, which can be described by its covariance matrix and mean vector [Richards and Jia, 1999]. An N-dimensional normal distribution is defined as:

$$p(\mathbf{x}) = [(2\pi)^{N/2} |C|^{1/2}]^{-1} \exp\{(-1/2)(\mathbf{x}-\mathbf{m})^t C^{-1}(\mathbf{x}-\mathbf{m})\} \quad (4.1)$$

where N is the number of dimensions (spectral bands) used.

$\mathbf{x}$ is the vector location in the N-dimensional pixel space.

C is the covariance matrix of the distribution.

$\mathbf{m}$ is the mean position of the spectral class.

The only input training parameters for this method is the maximum standard deviation from the mean, which has a default value of 3.0. Regions with bright pixels usually will use a lower standard deviation value, while dark pixels require higher values. A region with a standard deviation value too high will undesirably saturate the image with that region, and if the value is low there will be too many unclassified pixels. Figure 4.4 shows the segmented Hyperion image.

One of the drawbacks of the parallelepiped method is that it may fail to classify some of the pixels in the image, but for our cases that was not so much of a problem. The percentages of unclassified pixels were rather small and the results were much better than any of the other methods tested. These unclassified pixels had different statistical properties than the properties of the selected regions of interest and could be classified as separate classes. In fact, when two more regions of interest were added from the areas of

unclassified pixels, the segmentation was almost complete with very few unclassified pixels. The segmentation of the Landsat images, which will be discussed in the next chapter, classified more than 99% of the total number of pixels using only three regions (water, vegetation and urban).

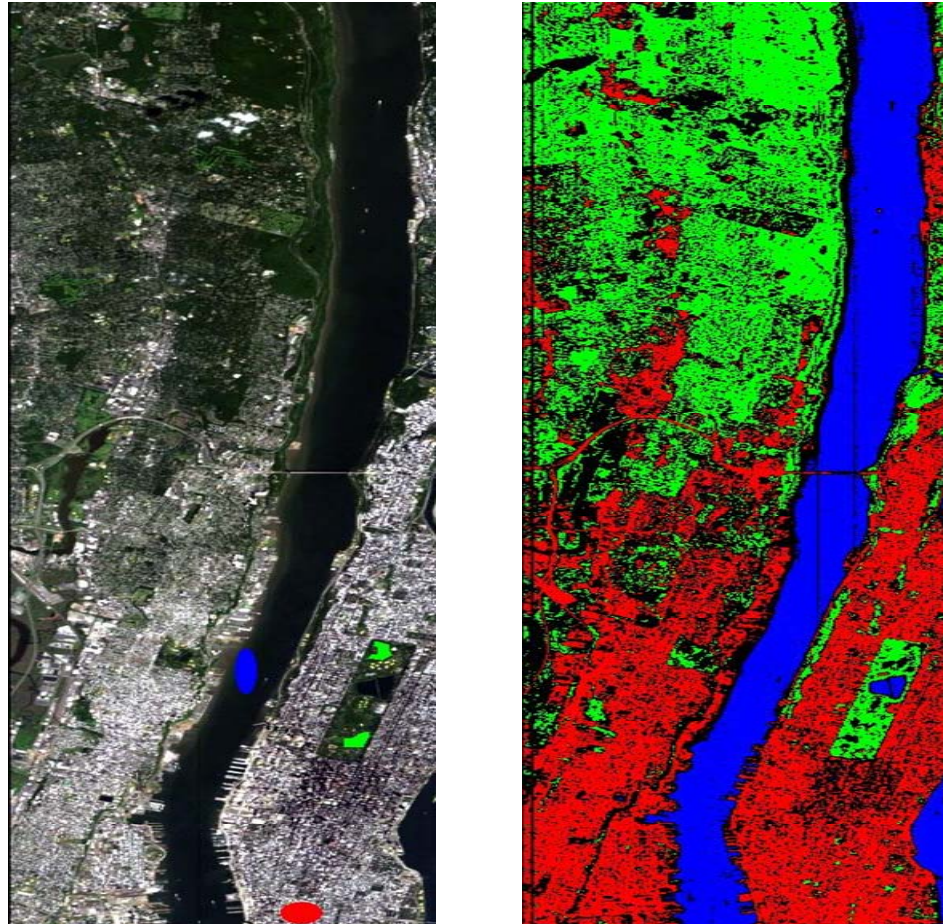


Figure 4.4: (Left) Original RGB image from Hyperion from which the three regions of interest were selected: water, vegetation and urban. (Right) Segmented image obtained by training the image using the selected regions of interest. The green (brightest) area shows the vegetation region (rectangle at the lower right corresponds to Central Park with its lake), the blue (darkest) area represents water features and the red (mid-gray) area represents the urban structures.

4.3 Hyperion Data Analysis

We performed the atmospheric correction using all visibilities (20 km to 120 km) and extracted the vegetation and urban regions from each file.

To determine the VIS-SWIR coefficient we obtained the linear fit of the scatter plot of the 2130 nm channel (x-axis) and the 470 nm (for C470) and 660 nm (for C660) channels. Figure 4.5 shows an example of these scatter plots with their corresponding slope. The slopes were calculated for the vegetation and urban regions (the urban region was initially called buildings region) for all visibility values.

The results of these correlation coefficients as a function of visibility are seen in Fig 4.6. We can see that from 80 km and above, the correlation coefficient for the VIS-SWIR ratio are quite stable. The TOA reflectance ratio of VIS-SWIR is approximately equal to the surface reflectance ratio of VIS-SWIR in this condition so the VIS/SWIR correlation coefficients can be calculated directly.

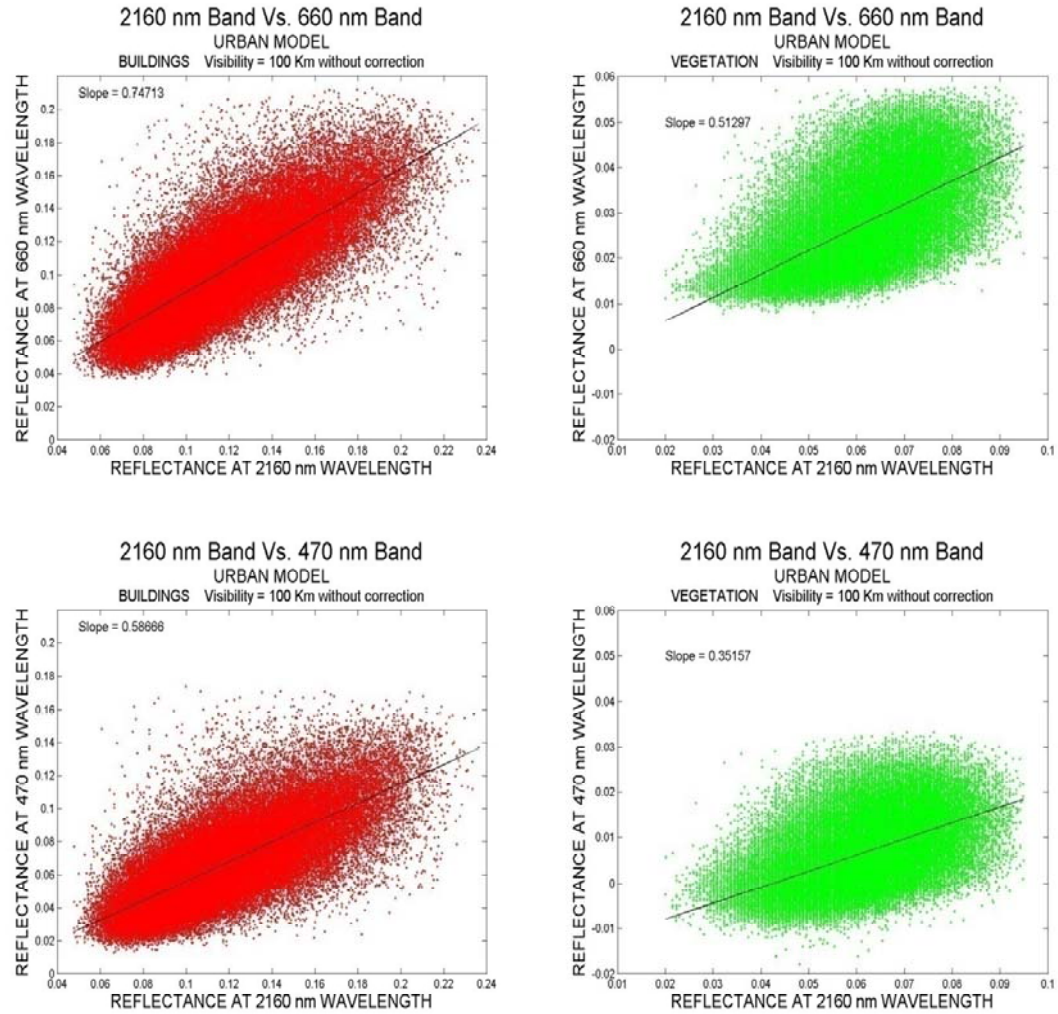


Figure 4.5 Scatter plots of the from the 2160nm and 660 nm channels (Top 2 panels) and the 2160nm and 470 nm channels (Bottom 2 panels) for the urban (on the left) and vegetation regions (on the right) from atmospherically corrected file using 100 km visibility.

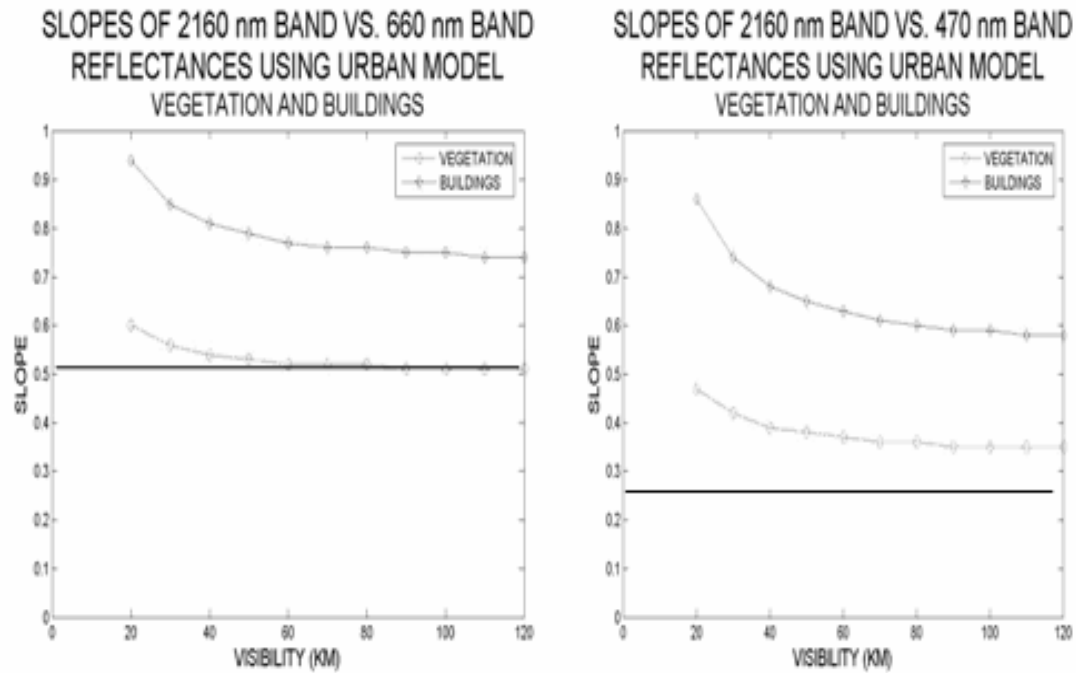


Figure 4.6 Mean slopes of the 2160 nm and 660 nm bands (left) and the 2160 nm and 470 nm bands (right) for the vegetation and urban (labeled as buildings) regions at each visibility. The lower bars correspond to the values of the MODIS C004 ATBD VIS-SWIR correlation coefficients of 0.5 for the 660 nm and 0.25 for the 470 nm band.

We can see that at high visibility, (low AOD, which was the case for this image), the results for the VIS-SWIR coefficient for the 660 nm channel (c660) from the vegetation region agrees with the 0.5 value previously assumed by MODIS, but it is much higher for the urban (buildings) region. For the 470 nm channel the coefficients (c470) are higher in both regions. There is a significant change when the atmosphere is assumed too thick (low visibility values).

As seen in chapter 3, the MODIS C005 algorithm incorporates the values of the Modified Vegetation Index into the calculations of the new VIS-SWIR coefficients. We wanted to verify if this parameter added any real improvement to the algorithm. We calculated the MVI for the vegetation and urban classes. Our results show that this

parameter can do a pretty good job of class separation. Therefore, its inclusion in the current MODIS algorithm proved to be advantageous.

The results from this single image cannot by itself be used to make any definite assumption of the appropriate coefficients to use over New York City, but it shows the need for further investigation. In the next chapter, we used Landsat imagery to further study these approaches. Land imagery has the same spatial resolution and there are more files available.

CHAPTER 5

Landsat Imagery Analysis

As shown in figure 4.1, the width of the swath of the Landsat imagery is 185 km, which allows a revisit time of 16 days. The release of Landsat Level 1B data in archived form opens the possibility to further test the robustness of the urban regions. Archived Landsat data can be obtained from the Earth Explorer website [Earth Explorer]. Most of the files are currently available for immediate download. If they are not readily available, they can be ordered and within several days, they will be processed and made accessible for download.

The Landsat 7(ETM+) sensor has six bands with 30 m spatial resolution (bands 1-5 and 7). Unlike Hyperion that had a narrow 10nm bandwidth for all bands, the bandwidths of the Landsat 7 are different for each band and much wider, ranging from 60 nm to 270 nm. The ranges for each band are (in nm): 450 to 520, 520 to 600, 630 to 690, 760 to 900, 1550 to 1750 and 2080 to 2350 for bands 1, 2, 3, 4, 5, and 7 respectively, as seen in table 5.1. The corresponding bandwidths are: 70, 80, 60, 140, 200 and 270, and the center (mean) wavelengths values are: 485, 560, 660, 830, 1650, and 2215. The Thermal and PAN bands are not used in this study

Satellite	Sensor	Bandwidths	Resolution	Satellite	Sensor	Bandwidths	Resolution
LANDSATs 1-2	RBV	(1) 0.48 to 0.57	80	LANDSATs 4-5	MSS	(4) 0.5 to 0.6	82
		(2) 0.58 to 0.68	80			(5) 0.6 to 0.7	82
		(3) 0.70 to 0.83	80			(6) 0.7 to 0.8	82
	MSS	(4) 0.5 to 0.6	79			(7) 0.8 to 1.1	82
		(5) 0.6 to 0.7	79		TM	(1) 0.45 to 0.52	30
		(6) 0.7 to 0.8	79			(2) 0.52 to 0.60	30
		(7) 0.8 to 1.1	79			(3) 0.63 to 0.69	30
LANDSAT 3	RBV	(1) 0.505 to 0.75	40			(4) 0.76 to 0.90	30
		(4) 0.5 to 0.6	79			(5) 1.55 to 1.75	30
	MSS	(5) 0.6 to 0.7	79			(6) 10.4 to 12.5	120
		(6) 0.7 to 0.8	79			(7) 2.08 to 2.35	30
		(7) 0.8 to 1.1	79	LANDSAT 7	ETM ⁺	(1) 0.45 to 0.52	30
		(8) 10.4 to 12.6	240			(2) 0.52 to 0.60	30
						(3) 0.63 to 0.69	30
						(4) 0.76 to 0.90	30
						(5) 1.55 to 1.75	30
						(6) 10.4 to 12.5	60
						(7) 2.08 to 2.35	30
						PAN 0.50 to 0.90	15

Table 5.1 Landsat satellites and sensors bandwidths and spatial resolution.

5.1 Landsat Data Collection and Preparation

The Landsat data comes in a compressed file. The uncompressed file produces a folder containing nine .TIF image files and three documentation file. The documentation files include a readme file and a metadata file. See the README.GTF file on appendix B and the metadata file on appendix C to see the information contained in these files.

Specific data was obtained by searching for Landsat 7 (ETM+) data that covered a 45 km x 45 km area that included New York City (Landsat Paths 013 and 014, Row 032).

The images were required to meet the following criteria:

- 1) No clouds over New York City and its surroundings (some images have clouds outside of the area of interest).
- 2) The images should have been acquired on days of low aerosol loading ($AOD_{550} < 0.1$) over New York City. One image with very high aerosol loading ($AOD_{550} > 0.5$) was also

used for assessment of aerosol retrieval capabilities for comparison. AOD information for the day and approximate time of the images acquisition were obtained from level 1.5 AOD data (Clouds cleared) from the AERONET site at The City College of New York [AERONET website].

The search for suitable data was greatly cumbersome due to a lack of availability. First, The Landsat 7 (ETM+) loss its scan line corrector (SLC) on May 31, 2003 [Landsat SLC-off website] so the images acquired since then are missing 22% of the data on most part of the image. Only the centers of the images have no missing data, but for Paths 013 and 014, these good coverage area lies too far outside of our area of interest. Second, The City College AERONET site does not have any data prior to 2001 (There is one measurement for March 2, 2000). Therefore my search was limited to 2001, 2002, and up to May 31, 2003 (for Landsat SLC ON images).

Figure 5.1 shows the selected Landsat images that met the search criteria. Figures 5.2, 5.3, and 5.4 show the AOD measurements from AERONET corresponding to the dates of acquisition of the Landsat images. Four of the images selected were acquired on days with very low aerosol loading ($AOD_{500} < 0.1$) (acquired on February 19, 2002; September 8, 2002; September 24, 2002; and January 14, 2003). The other image was acquired a day with a very high aerosol loading ($AOD_{500} = 0.98$) (acquired on August 14, 2002). Table 5.2 shows some information about these files.

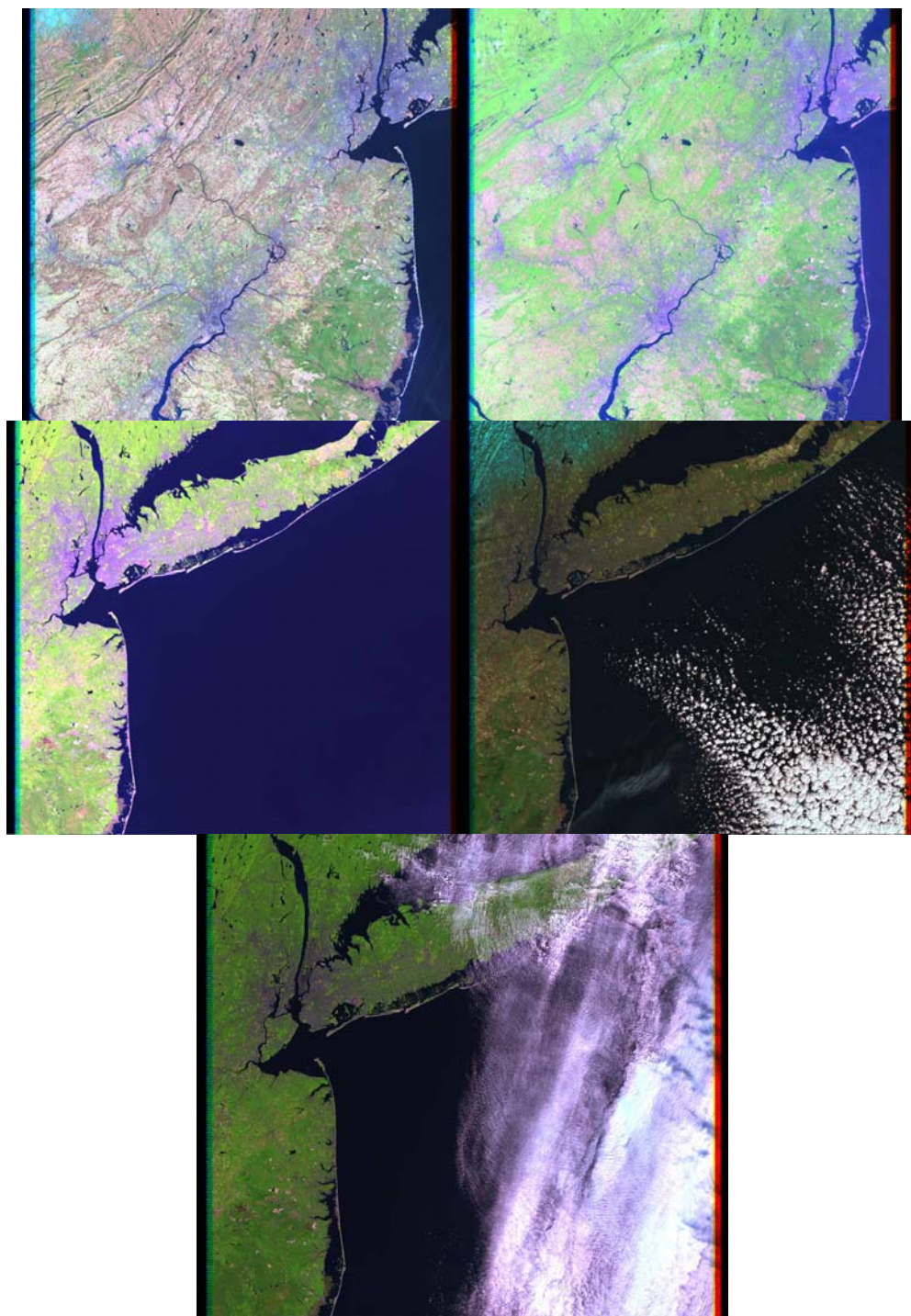


Figure 5.1: Landsat images over New York City acquired between February 19, 2002 and January 14, 2003.

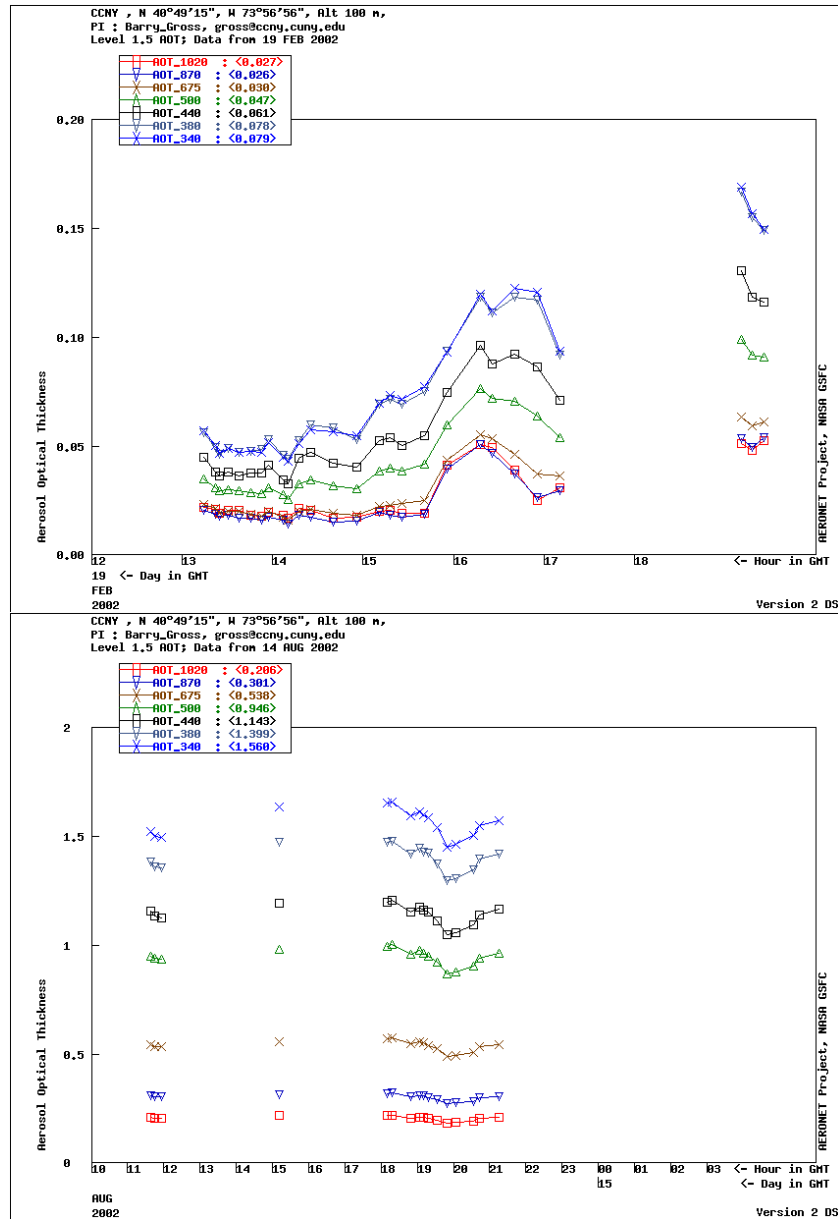


Figure 5.2: AOD data acquired on February 19, 2002 (Top) and August 14, 2002 (Bottom) from the Aeronet site at The City College of New York in New York City.

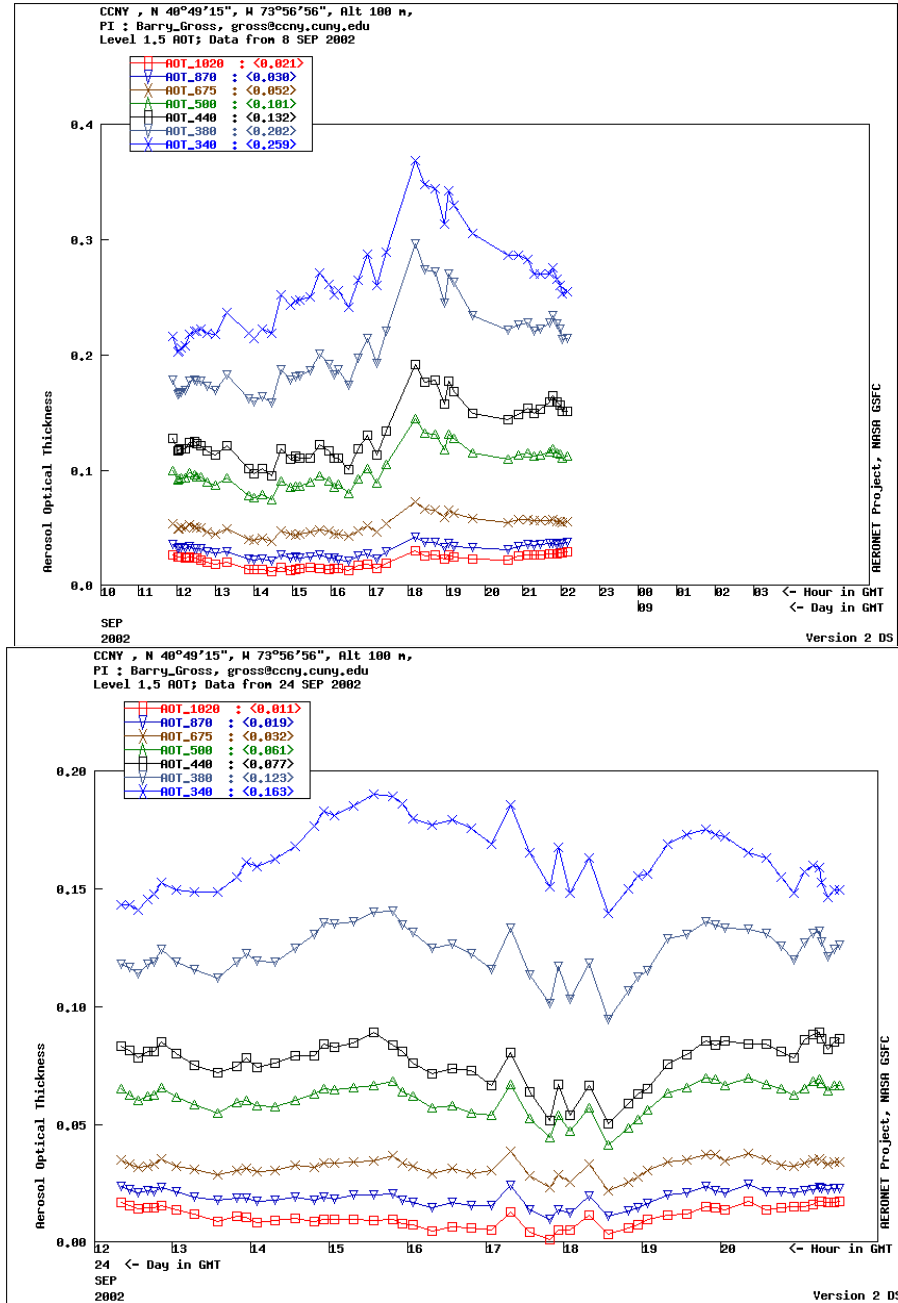


Figure 5.3: AOD data acquired on September 8, 2002 (Top) and September 24, 2002 (Bottom) from the Aeronet site at The City College of New York in New York City.

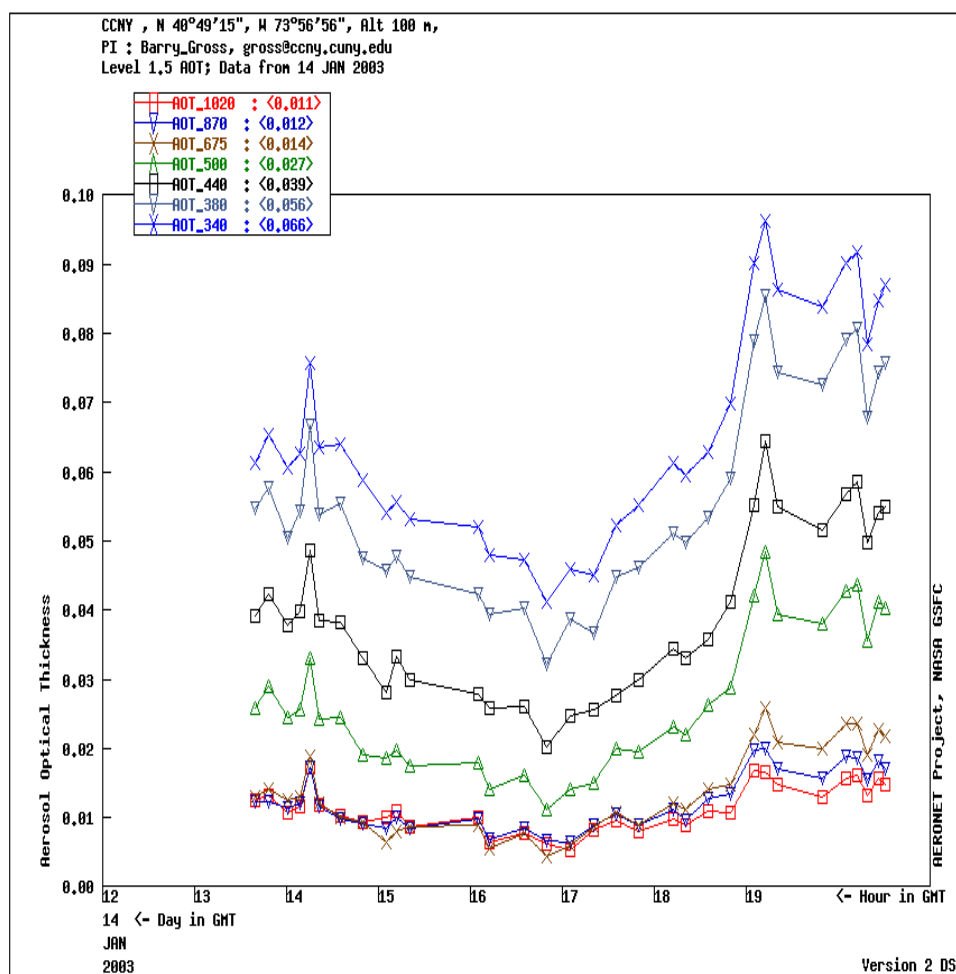


Figure 5.4: AOD data acquired on January 14, 2003 from the Aeronet site at The City College of New York in New York City.

Image Acquisition date	Image Acquisition time (GMT)	Zenith angle	AOD500	AOD time (GMT)	Average AOD500 for the day
02/19/02	15:28	56.94°	0.038	15:20	0.047
08/14/02	15:28	33.52°	0.98	15:11	0.946
09/08/02	15:21	40.16°	0.089	15:27	0.101
09/24/02	15:21	45.18°	0.066	15:19	0.061
01/14/03	15:22	65.91°	0.018	15:20	0.027

Table 5.2 Information for Landsat7 ETM+ images used for analysis. The AOD500 time (GMT) is the time the AOD measurement was taken, closest to the time of acquisition of the Landsat image.

As stated at the beginning of this section, the Landsat images were obtained as individual GeoTiff images (one image per band). The following section details the procedure to convert them into radiance files and put them into a format to be used for the atmospheric correction software.

5.2 Conversion of Landsat GeoTIFF Image Files into one Multilayered Radiance File using ENVI Software

This section details the steps to convert Geographic Tagged Image File Format (GeoTIFF) image files from the Landsat 7 ETM+ sensor into one radiance image file using the ENVI software. The individual grayscale GeoTIFF images were converted into radiance values and were put together into layered data files containing the bands 1, 2, 3, 4, 5, and 7 (30 m spatial resolution bands). The layered radiance files were also converted into a BIL format to be used with the ENVI atmospheric correction tool (FLAASH). We used the ENVI 4.1 version of the software, but newer versions also contain the tools necessary to perform these procedures (most of the instructions for the tools used in these

procedures were also found on the User's Guide of the latest ENVI software). The conversion from grayscale (0-255) values to floating point radiance is detailed on Chapter 11 - Data Products, of the Landsat 7 Science Data Users Handbook [Landsat handbook] and is given as:

$$L_{\lambda} = ((LMAX_{\lambda} - LMIN_{\lambda}) / (QCALMAX - QCALMIN)) * (QCAL - QCALMIN) + LMIN_{\lambda} \quad (4.1)$$

The Science handbook contains the tables with all these values (except QCAL, which is the grayscale value of the pixel to be converted to radiance), but *only* the values from the metadata files should be used since they contain the actual values used for that particular file. For example, for the first file, the detectors for bands 1, 2, 3, 4, 5, and 7 were all set to High Gain, but we also found other cases where band 4 was set to Low Gain, so we had to update the band math equation.

The procedure for the conversion is as follow:

- 1) Open the ENVI software.
- 2) Load all six .TIF image files:
 - From the ENVI's toolbar, click on 'File', then on 'Open Image File', which will open a file browser window.
 - Select the folder that contains the .TIF files to be processed. Hold down the 'CTRL' button on the keyboard and select the image files, then press the 'open' button. For our purposes the following files were selected for processing:

'Filename' _B10.TIF	(image file for Band 1)
'Filename' _B20.TIF	(image file for Band 2)
'Filename' _B30.TIF	(image file for Band 3)
'Filename' _B40.TIF	(image file for Band 4)
'Filename' _B50.TIF	(image file for Band 5)
'Filename' _B70.TIF	(image file for Band 7)

(Where 'Filename' corresponds to the name of the Landsat file being processed).

- The files will load and the 'Available Bands List' window will pop on the screen with all the loaded bands listed.
- Click (inside the 'Available Bands List' window) on the 'Band 1' under the name of the band you want to load to display one of the bands (this is optional but it is a good idea, in order to make sure that you have the right image). If you want to display the image as RGB, click on the 'RGB Color' button and click on the 'Band 1' names for bands 3, 2, and 1 in that exact sequence (for Landsat data, band 3 is the Red band, band 2 is the Green band and band 1 is the Blue band), then click on the 'Load RGB' button to display the image.

3) Select 'Basic Tools' on the ENVI toolbar, then select 'Band Math' from the pull down menu. The 'Band Math' window will appear. You need to enter the equation for each band individually. It is best to write all the formulas in a text file with extension .exp that you can load into the dialog box (use Notepad or WordPad, because ENVI does not accept the format of Microsoft Word files). The heading 'ENVI EXPRESSIONS' must be included (otherwise you will get a file reading error). The following is an example of how to write the equations for radiance conversion (using eq. 4-1):

a) For band 4 set to High Gain:

ENVI EXPRESSIONS

$((191.6+6.2)/254.0)*(\text{float}(b1)-1.0)-6.2$

$((196.5+6.4)/254.0)*(\text{float}(b2)-1.0)-6.4$

$((152.9+5.0)/254.0)*(\text{float}(b3)-1.0)-5.0$

$((157.4+5.1)/254.0)*(\text{float}(b4)-1.0)-5.1$

$((31.06+1.0)/254.0)*(\text{float}(b5)-1.0)-1.0$

$((10.80+0.35)/254.0)*(\text{float}(b6)-1.0)-0.35$

b) For files where band 4 is set to Low Gain replace the equation for b4 with:

$$((241.1+5.1)/254.0)*(\text{float}(b4)-1.0)-5.1$$

- In the above expressions b1, b2, b3, b4, b5, and b6 are the variables representing bands 1, 2, 3, 4, 5 and 7 respectively. When writing a math expression to be used with these tools, the names for all variables must begin with the letter 'b' or 'B' followed by up to 5 digits.
 - Once the math expressions are loaded into the window under the name 'Previous Band Math Expressions:' click on the expression for the band to be processed. The expression will be highlighted and it will appear in the window under 'Enter an expression:'. Click OK, the 'Variables to Bands Pairings' window will appear to define to variable.
 - Click on the 'Band 1' under the name of the band being processed and enter the name of the output file (click the 'Choose' button to browse to the folder where you want to store the radiance files then enter the name for the radiance file like for example: 'Radiance_Band1') and click OK. The grayscale image will be converted into a radiance data file and saved in the chosen folder. The processed files will also appear on the 'Available Bands List' window from which they can be selected to be displayed.
- 4) To put the six bands together in a single layered file click on 'Basic Tools' on the ENVI toolbar, then select 'Layer Stacking' from the pull down menu. The 'Layer Stacking Parameters' window will appear.
- Click on the 'Import File...' button, the 'Layer Stacking Input File' window will appear. If this step is done right after the band math operation (step 3) all radiance

files should appear in the ‘Select Input File:’ section, click and drag the mouse to select all bands at once. If the radiance files did not appear, they will have to be loaded by clicking the ‘Open’ button, then selecting ‘New File...’ from the pull down menu, and browsing for the radiance files.

- Click the OK button on the ‘Layer Stacking Input File’ window, which will return you to the ‘Layer Stacking Parameters’ window.
 - Warning: If the radiance files are not listed in order from band 1 to band 6 (band1 must be on top, followed by band 2, and so on), click on the ‘Reorder Files...’ button. The ‘Reorder Files’ window will pop with all files in it. Click and drag the files into their proper sequence and click OK. This is very important since these will be the order of the stacking.
 - Enter the name of the layered radiance file, and verify the information on the right side regarding the map projection to see if it was properly loaded (this information should match the information in the metadata file) and click OK to begin the layering process. Once the process is finished the file will be saved in the chosen folder and it will also appear on the ‘Available Bands List’, from where they can be selected to be displayed.
- 5) To convert the file into a BIL format, which is the format used by FLAASH the atmospheric correction tool (FLAASH also works with BIP files).
- Click on ‘Basic Tools’ on the ENVI toolbar, then select ‘Convert Data (BSQ, BIL, BIP)’ from the pull down menu, the ‘Convert File Input File’ window will appear.

- Click on the layered radiance file, do any spatial subset if so desired, and click OK, the window will close and the 'Convert File Parameters' window will appear.
- Click on the BIL (or BIP) button, enter the filename (click the 'Choose' button to select the folder where you want the file to be saved) and click OK. Once the process is finished the file will be saved in the chosen folder and it will also appear on the 'Available Bands List', from where it can be displayed.

5.3 Landsat Data preparation

After the conversion to radiance, we selected a 1500 by 1500 pixels (45km x 45 km) area encompassing New York City, as seen in figure 5.5. All Landsat files were subset using the same map coordinates to ensure the selection of the same area of interest.

Just as we did with the Hyperion image, we performed the atmospheric correction of the radiance files using the FLAASH tool from ENVI. For the selection of models parameters, we chose the US Standard atmospheric model and the urban aerosol model. We ran the corrections using several values of visibilities. For the images with low aerosol loading we used a range from 20 to 120 Km in 20 km increments. For the high AOD image, we also used 1, 5, 10, and 15 km, in addition to the 20 to 120 km visibilities values used for the low AOD files. For each of the atmospherically corrected image file, we saved all six Landsat bands in separate files in ASCII format. We removed the ENVI heading included on each of these files and save them as data files which could be loaded directly into MATLAB.

The images were segmented into three regions: water, vegetation and urban. The segmentation process was detailed in the Hyperion chapter. Figure 5.6 shows an example a segmented Landsat image (cropped to the New York City area).

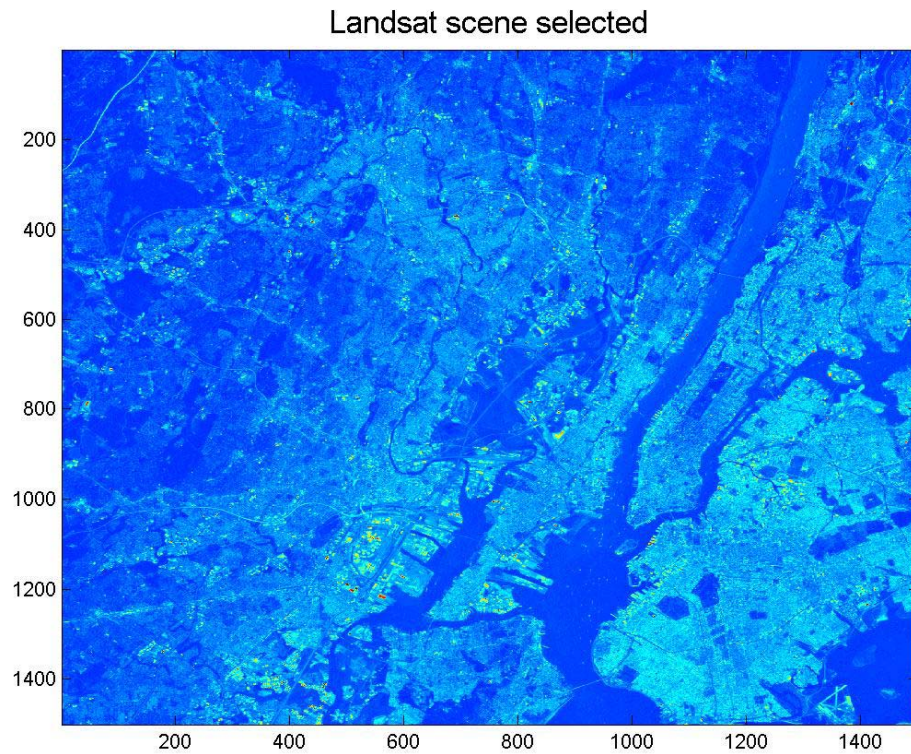


Figure 5.5: View of a 45 km x 45 km (1500 x 1500 pixels) image over New York City from Landsat.

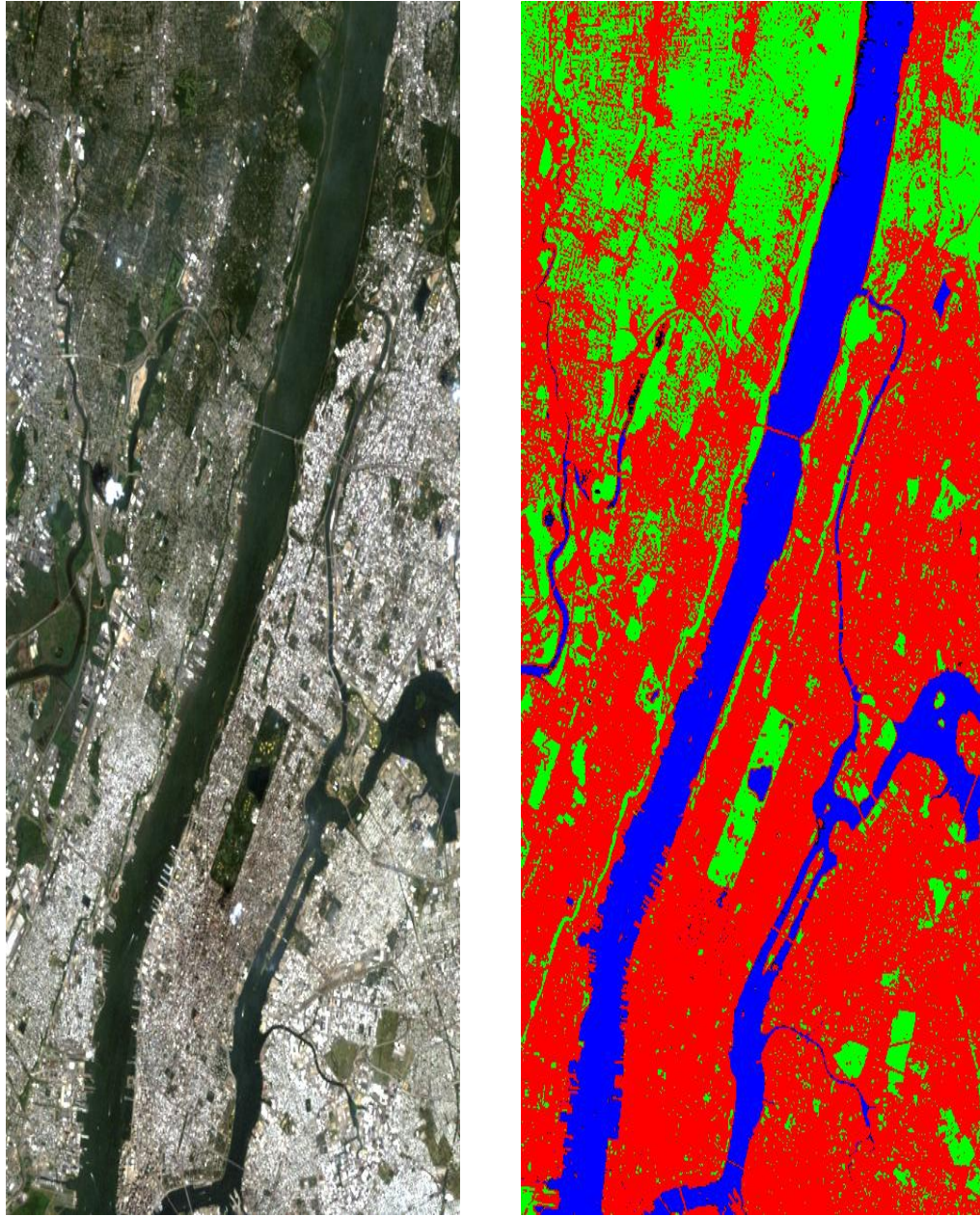


Figure 5.6: Landsat image over New York City. (Left) RGB image. (Right) Segmented image into water (blue/darkest), vegetation (green/brightest) and urban (red/gray) regions.

5.4 Landsat Data Analysis and Results

For the Landsat imagery we used the values of 485 nm and 470 nm interchangeably for VIS wavelength values because they are represented by the same band (band 1). Similarly we use the values of 2130, 2200, 2215 for the SWIR wavelength (they are all included in band 7). Table 5.1 shows these ranges of values. 485 nm and 2215 nm are the center wavelengths of Landsat's bands 1 and 7 respectively.

NOTE: Band 6 of the Landsat 7 (ETM+) sensor is the Thermal band, which was not used in this research, therefore, the Band 6 that we refer to in this thesis actually correspond to Landsat's Band 7.

For our analysis, we divided the 1500 x 1500 pixels (45 km x 45 km) data set into 9 segments of 500 x 500 pixels (15 km x 15 km). For each segment, we calculated the ratios between the VIS (485nm blue and 660nm red bands) and MIR (2215nm band), as well as the mean and standard deviation. The ratios were obtained by applying a linear regression to the VIS-MIR scatter plot. We performed these procedures for the vegetation and urban region for all atmospherically corrected files. Prior to the calculation the data was cleared of outliers using the interquartile range (IQR) boundaries. We ignored all points that lied outside the range of $1.5 \cdot \text{IQR}$ above the upper quartile (upper bound), and $1.5 \cdot \text{IQR}$ below the lower quartile (lower bound).

Figure 5.7 shows the VIS-MIR scatter plots of the urban region for 20 km and 120 km visibility values for the atmospheric correction for the February 19, 2002 image. For this day, the AOD at 500nm was .038 allowing us to get very good observations of the surface. Fig 5.8 shows the same for the vegetation region. The main focus of this analysis is to verify that for vegetation pixels, the spectral ratios are close to C004 values

at high visibility while the urban scene is much higher and in good agreement with previous estimations of the VIS-SWIR from the previous datasets we used in this section.

Figures 5.9 and 5.10 shows the results of the VIS-MIR slopes (correlation coefficients) for all visibilities used for the atmospheric correction for band 1 (470nm) and band 3 (660 nm) respectively. We note in particular that for the 660nm channel, the slopes are quite consistent with all previous results and are fairly invariant. However, for the summer observations, the 470 nm correlation slopes are much higher than those in the winter. This might be a function of the fact that the solar zenith angles are much less and we might be seeing enhanced backscatter (i.e. the hotspot effect) which occurs in the backscatter direction.

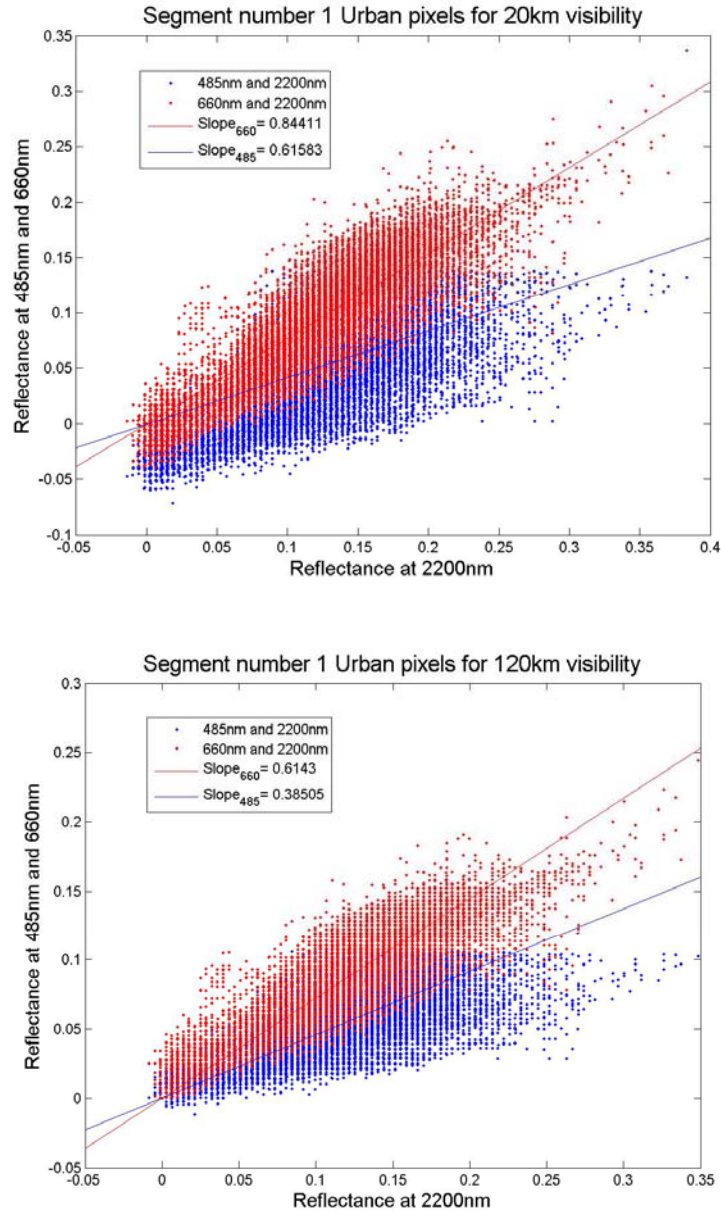


Figure 5.7 Scatter plots of VIS-MIR bands of urban pixels using (Top) 20 km visibility and (Bottom) 120 km visibility values for the atmospheric correction. The x-axis values correspond to the reflectance at 2200nm wavelength and the y-axis values correspond to the reflectances at 485 and 660nm wavelengths. The darker dots correspond to the 485nm values and the lighter dots correspond to the 660nm values. The straight lines represent the linear fit for each cluster.

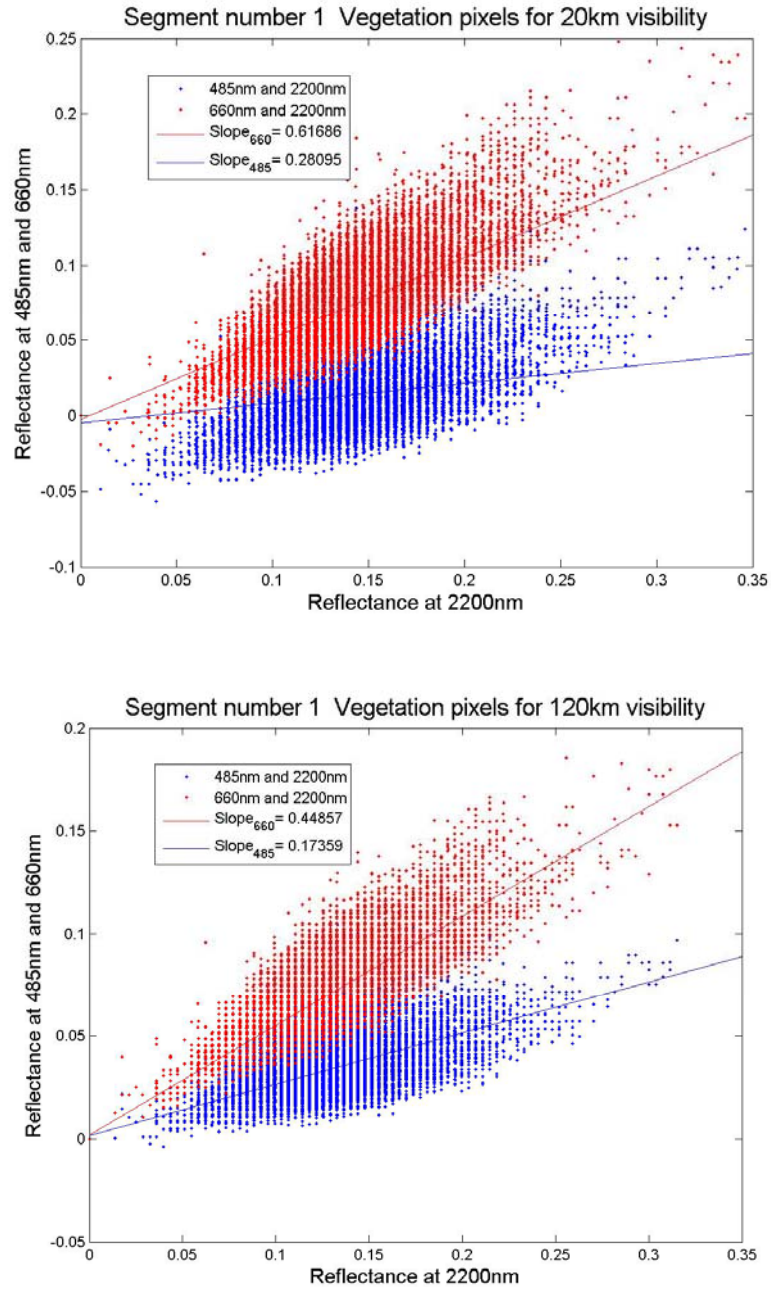


Figure 5.8 Scatter plots of VIS-MIR bands of vegetation pixels using (Top) 20 km visibility and (Bottom) 120 km visibility values for the atmospheric correction. The x-axis values correspond to the reflectance at 2200nm wavelength and the y-axis values correspond to the reflectances at 485 and 660nm wavelengths. The darker dots correspond to the 485nm values and the lighter dots correspond to the 660nm values. The straight lines represent the linear fit for each cluster.

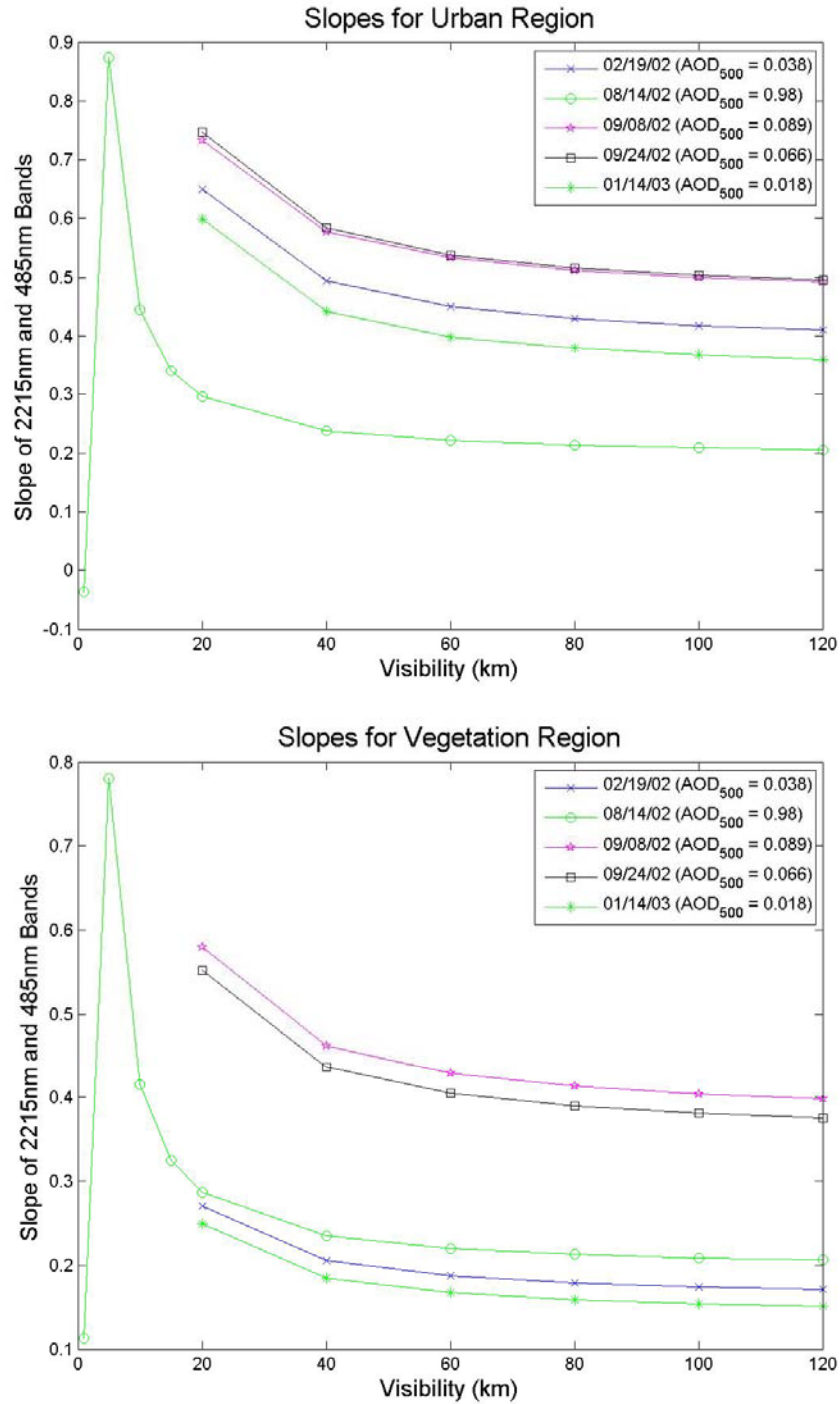


Figure 5.9: Values of slopes of 485nm band and 2215nm band for urban region (Top) and vegetation region (Bottom) from Landsat data acquired on different dates with different aerosol loading. The data was atmospherically corrected using visibilities of 20 km to 120 km for low AOD data, and 1 km to 120 km for high AOD.

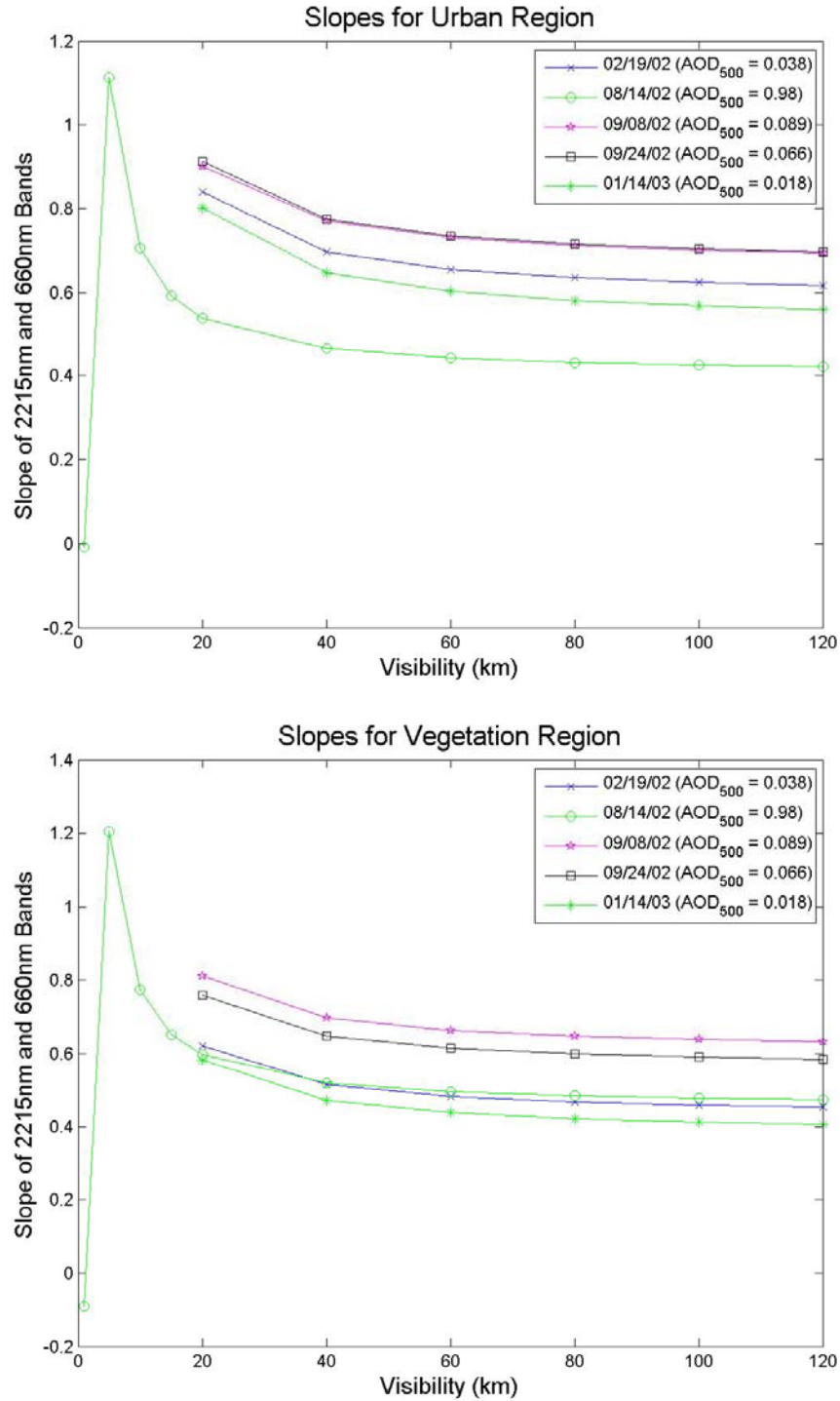


Figure 5.10: Values of slopes of 660nm band and 2215nm band for urban region (Top) and vegetation region (Bottom) from Landsat data acquired on different dates with different aerosol loading. The data was atmospherically corrected using visibilities of 20 km to 120 km for low AOD data, and 1 km to 120 km for high AOD.

These results showed good agreement of the VIS-SWIR ratios for summer months. Even for the case with high AOD, there is a good agreement when the correct visibility value is used (about 10 km in this case). In winter months, there is poorer agreement. We believed that this is likely due to changes in vegetation. Further studies need to be done to investigate these anomalies.

Another approach is to use the Landsat data and the radiative transfer analysis to determine the most suitable coefficients for New York City. This analysis is done in the following chapter.

CHAPTER 6

Current Work: AOD Retrieval Using Landsat with Land Classification Spectral Ratios

This chapter covers the work on Landsat data using radiative transfer analysis that is currently being done at the Remote Sensing Lab at The City College of New York. The generation of the radiative transfer reflectances is being currently implemented by Dr. Bomidi Madhavan, a postdoctoral fellow at this lab. This work is on its preliminary stage but some of the results so far have been encouraging. First we will provide a brief introduction to aerosol retrieval and the radiative transfer analysis, followed by some of the results obtained using this approach.

The retrieval of aerosol optical depth (AOD) over land remains difficult and demanding because the surface reflectance is generally higher, so it provides less sensitivity to changes in aerosols, and also varies spatially and temporally. Inaccuracies of 0.01 of the surface reflectance can result in AOD variations of 0.1 [*Hauser et al.*, 2004]. The surface reflectance of land surface is usually unknown and varies with wavelength. Several algorithms have been applied to satellite databases to solve the inverse problem of separating the surface and atmospheric scattering contributions. As previously stated, MODIS retrievals of aerosol over land are based on the correlation of surface reflectance in the visible and shortwave infrared (SWIR) [*Kaufman et al.*, 1997; *Remer et al.*, 2005]. Dark, dense vegetation has a very low reflectance in the blue and red

wavelength regions and the surface reflectance at wavelengths in the visible are proportional to those in the mid-infrared [*Hansen et al.*, 1990].

On the other hand, with the high resolution of Landsat, we should be able to attempt retrieval of AOD on spatial scales where we can isolate the difference between vegetation and urban surfaces. In particular, it is interesting to explore the retrieval over domains of vegetation and urban that were very close and therefore shared the same aerosol loading. This would provide a good validation of the AOD.

6.1 Aerosol Optical Depth Retrieval Algorithm

Our retrieval technique consists of three steps. The first step is to consider a bimodal aerosol size distribution and obtain the aerosol optical properties using the microphysical parameters in Lorentz-Mie theory. The second step involves computing top of the atmosphere (TOA) reflectance from the irradiance measurements obtained by incorporating the previously derived aerosols properties in the Atmospheric Radiative Transfer model. Finally, look up tables (LUTs) of TOA reflectance were generated with respect to aerosol optical depth (AOD) and surface albedo and were used to retrieve the visible AOD from the LANDSAT observed TOA reflectance values for different locations at 470 nm band (Band 1) and 660 nm band (Band 3) with respect to 2200 nm band (Band 6).

6.1.1 Aerosol optical properties for a bimodal size distribution using Mie Theory

In this study, a more realistic bimodal aerosol size distribution was adopted with two distinct classes: fine and coarse particles, each having a distinct mode in the aerosol

volume distribution function. We assume that the aerosol size distribution can be described by a combination of two lognormal distributions that represent fine (f) and coarse (c) modes [Dubovik *et al.*, 2002]:

$$n(r) = \frac{dV(r)}{d \ln r} = \frac{N_f}{\sqrt{2\pi}\sigma_f} \exp\left(-\frac{[\ln r - \ln R_f]^2}{2\sigma_f^2}\right) + \frac{N_c}{\sqrt{2\pi}\sigma_c} \exp\left(-\frac{[\ln r - \ln R_c]^2}{2\sigma_c^2}\right) \quad (6.1)$$

where r (in μm) is particle radius, N (in cm^{-3}), R (in μm) and σ^2 are the volume concentration, volume median particle radius and variance, respectively. The particle number size distribution is linked with volume size distribution. Here the shape of the aerosol size distribution is assumed to be independent of altitude so that the vertically integrated number density N is related by a scale height factor to the surface number density. Table 6.1 shows the size distribution parameters obtained by measurements from the CCNY AERONET site on August 14, 2002, which was a high AOD day, thus parameters provided were more accurate.

The variables physically independent of particle size (if we do not consider processes like coagulation) are the total number of particles and the particle number

$$\text{fraction} \quad \hat{v}_f = \frac{N_f}{N_f + N_c} \quad (6.2)$$

where N_f and N_c are column numbers of fine and coarse mode particles, and $\hat{v}_f$ is a parameter of the normalized total aerosol size distribution:

$$n(r) = \hat{v}_f n_f(r) + (1 - \hat{v}_f) n_c(r) \quad (6.3)$$

Where $n_f(r)$ and $n_c(r)$ are (normalized) size distribution functions for respective fine and coarse mode aerosol. Table 6.1 summarizes the microphysical parameters of aerosol bimodal size distribution in our present study. These values are the true values observed on 081402 (high AOD day) at the CCNY AERONET site.

Size distribution parameter	Fine mode (f)	Coarse mode (c)
Mean particle radius (r_m)	$r_f = 0.11756 \mu m$	$r_c = 1.29038 \mu m$
Standard deviation (σ)	1.4918	1.7683
Number concentration (N_i)	9.9949	0.0010
Fine mode fraction ($\hat{v}_f$)	0.99989	

Table 6.1 Microphysical parameters of assumed aerosol bimodal size distribution.

Light extinction by small spherical particles can be described by Mie theory [Bohren and Huffman, 1983; Mischchenko, 2002], with key parameter as the complex refractive index of the particle relative to the surrounding air: $m_\lambda = n_\lambda + ik_\lambda$. We can relate (through the Lorentz-Mie theory) the aerosol optical depth τ_a with size distribution and complex refractive index as

$$\tau_\lambda = \int_{r_{\min}}^{r_{\max}} n(r) Q_{\text{ext},\lambda}(r, m_\lambda) dr \quad (6.4)$$

$$\tau_\lambda = \int_{r_{\min}}^{r_{\max}} n(r) Q_{\text{ext},\lambda}(r, m_\lambda) dr \quad (6.5)$$

where $n(r)$ is particle number size distribution, and $Q_{\text{ext},\lambda}$ is extinction cross-sections.

Complex refractive index characterizes scattering (n_λ) and absorbing (k_λ) properties of the particles. For simplicity, the absorption and scattering effects of a soot-based nucleation mode were neglected (as were relative humidity induced variations of the real part of the refractive index) in favor of a fixed purely scattering refractive index (n_λ) of 1.5 for fine and coarse modes. The Mie derived parameters for lognormal distribution include extinction coefficient (β_{ext}), scattering coefficient (β_{sca}), single

scattering albedo (ω_0), asymmetry parameter (g) and the phase functions $p(\theta)$ for the scattering angles (θ) between 0 and π . These parameters are obtained for fine and coarse modes distinctly at 470 nm, 550 nm, and 660 nm wavelengths.

As per our assumption of the existence of distinct fine and coarse particle modes, the aerosol scattering phase function for the total aerosol size distribution can be written as the weighted mean:

$$P(\theta) = \frac{\beta_{ext}^f \omega_0^f P^f(\theta) + \beta_{ext}^c \omega_0^c P^c(\theta)}{\beta_{ext}^f \omega_0^f + \beta_{ext}^c \omega_0^c} \quad (6.6)$$

Similarly, the weighted mean of the single scattering albedo (ω_0) is given by:

$$\omega_0 = \frac{\beta_{ext}^f \omega_0^f + \beta_{ext}^c \omega_0^c}{\beta_{ext}^f + \beta_{ext}^c} \quad (6.7)$$

The superscripts f and c on the corresponding symbols indicate the ‘fine’ and ‘coarse’ mode components respectively. The total aerosol extinction can be expressed as a linear sum of the extinction coefficients of the fine (f) and coarse (c) modes:

$$\beta_{\lambda}^{ext} = \beta_{\lambda}^{ext,f} + \beta_{\lambda}^{ext,c} \quad (6.8)$$

The total aerosol extinction values (km^{-1}) at 470 nm and 660 nm are normalized with respect to that at 550 nm. The normalization of AODs at different wavelengths is

given by the relation: $\tau_{\lambda} = \tau_{550} \left(\frac{\beta_{\lambda}^{ext}}{\beta_{550}^{ext}} \right)$ (6.9)

where β_{λ}^{ext} is the total extinction coefficient due to fine and coarse modes observed, and τ_{λ} is the AOD at wavelength λ .

It is standard fare to expand the phase functions in terms of Legendre polynomials given by:

$$P_l(\xi) = \sum_{n=0}^N a_n p_n(\xi) \quad (6.10)$$

where $P_l(\xi)$ is the phase function, ξ is the cosine of the scattering angle θ , a_n are the Legendre coefficients and p_n are Legendre polynomials. Note that n represents the polynomial order of $P_n(\xi)$ and hence, the equation approximates the phase function as a polynomial of order N .

The Legendre polynomials are computed using the orthogonality of Legendre polynomials [Liou, 1992]:

$$a_n = \frac{(2n+1)}{2} \int_{-1}^1 P_l(\xi) p_n(\xi) d\xi \quad (6.11)$$

where the first term in the expansion gives the single scattering albedo [Liou, 1992]:

$$a_0 = \frac{1}{4\pi} \int_0^{2\pi} \int_{-1}^1 P_l(\xi) d\xi d\phi \quad (6.12)$$

For pure scattering ($\omega_0 = 1$), this satisfies the normalization condition: $a_0 = \omega_0 = 1$. It is customary to limit the number of Legendre polynomials in radiative transfer calculations to the number of computational polar angles, or streams, with the rationale being that little can be gained from phase functions that are more precise than the model that employs them [Thomas and Stamnes, 1999; p194]. Since the computation time is proportional to the cube of the number of streams [Stamnes et al., 2000], it is important to limit the number of streams (and hence the number of Legendre moments). This poses a problem for calculations involving the highly anisotropic phase functions associated with aerosols, as low-order Legendre polynomials are not capable of capturing the strong forward scattering peak. Modern radiative transfer codes (such as DISORT) account for

strong forward scattering by approximating the forward peak as a Dirac δ -function weighted according to the number of streams [Thomas and Stamnes, 1999, p 191; Wiscombe, 1977; Stamnes *et al.*, 2000]. The realistic phase functions are used for the present study. When realistic phase functions are not available, it is usual practice to derive using the Henyey Greenstein approximation computed by the asymmetry parameter (g). Note that HG phase functions are not accurate in backward scattering direction. The remainder of the phase function (i.e., polar angles outside of the forward peak) is scaled and the number of moments is limited to the number of streams. Accurate representation of the Legendre coefficients is necessary to avoid a nonsensical phase function expansion.

6.1.2 Atmospheric Radiative transfer calculations using SBDART model

Santa Barbara DISORT Atmospheric Radiative Transfer (SBDART) is a software tool that computes the plane parallel radiative transfer in clear and cloudy conditions within the Earth's atmosphere and at the surface [Ricchiazzi *et al.*, 1998]. In SBDART calculations, a user has freedom to set various parameters for the input to the model. The only subjectively selected parameters for the input are atmospheric profile (US 62), solar zenith angle, wavelength of observation, surface albedo feature, and aerosol properties which include aerosol optical depth, single scattering albedo and the computed Legendre moments of the phase functions for different wavelengths. The extinction optical depth for each layer is the sum of the layer optical depths due to molecular absorption, aerosols, clouds, and Rayleigh scattering. Since the gas transmission at a particular wavelength is itself broken into a sum of three exponential terms, the complete

solution to the multi-scattering radiative transfer involves three separate invocations of the DISORT algorithms to treat each of the k-distribution components. Each sub-calculation uses the same optical properties of clouds, aerosols, and Rayleigh scattering but different gas optical depths. A fraction of the solar input is assigned to each of the three k-distribution terms, with all the fractions summing to one.

The computation of input parameters by SBDART yields both the upwelling and down-welling irradiance (in $Wm^{-2}\mu m^{-1}$) at the top of the atmosphere (TOA) and at the surface for the computed solar zenith angles. The TOA reflectance is given by the ratio of upwelling diffuse radiance (in $Wm^{-2}\mu m^{-1}sr^{-1}$) to down-welling direct irradiance (in $Wm^{-2}\mu m^{-1}$). In our present study, we consider the following two classes, urban and vegetation. The values of the VIS-SWIR coefficients are shown in Table 6.2.

Parameter	WAVELENGTH		
	2130 nm (Band 6)	470 nm (Band 1)	660 nm (Band 3)
Surface Albedo (ρ_g)	$\rho_g^{2130}=[0:0.02:0.2]$	<u>URBAN</u>	<u>URBAN</u>
		$\rho_{g,470}^U=0.38 \rho_g^{2130}$	$\rho_{g,660}^U=0.69 \rho_g^{2130}$
		<u>VEGETATION</u>	<u>VEGETATION</u>
		$\rho_{g,470}^V=0.25 \rho_g^{2130}$	$\rho_{g,660}^V=0.5 \rho_g^{2130}$

Table 6.2 Best surface albedo spectral slope estimates for different channels and variation of the 2120 surface reflection.

The look up tables are generated for the TOA reflectance (*unitless*), namely $\rho_{TOA,U}^{470}(\theta_s, \tau_{550}, \rho_g^{2130})$, $\rho_{TOA,U}^{660}(\theta_s, \tau_{550}, \rho_g^{2130})$, $\rho_{TOA,V}^{470}(\theta_s, \tau_{550}, \rho_g^{2130})$ and $\rho_{TOA,V}^{660}(\theta_s, \tau_{550}, \rho_g^{2130})$ for different surface albedo values i.e., $\rho_g^{2130}=[0:0.02:0.2]$ at $\tau_{550} = [0:0.001:1.0]$ for the

urban and vegetation classes using model computed radiance values at different solar zenith angles.

6.2 Aerosol optical depth retrieval from LANDSAT datasets

As previously stated, in remotely sensed data analysis the retrieval of the surface reflectance is very important when we want to obtain atmospheric reflectance or atmospheric correction. The Landsat 7 system offers the unique capability to seasonally monitor important small-scale processes on a global scale, such as the inter- and intra-annual cycles of vegetation growth; deforestation; agricultural land use; erosion and other forms of land degradation; snow accumulation and melt and the associated fresh-water reservoir replenishment; and urbanization.

The Enhanced Thematic Mapper Plus (ETM+) instrument on the Landsat 7 spacecraft contains sensors to detect earth scene radiation in three specific ranges:

- **Visible and near infrared (VIS and NIR) bands** - bands 1, 2, 3, 4, and 8 (PAN) with a spectral range between 0.4 and 1.0 μm .
- **Short wavelength infrared (SWIR) bands** - bands 5 and 7 with a spectral range between 1.0 and 3.0 μm .
- **Thermal long wavelength infrared (LWIR) band** - band 6 with a spectral range between 8.0 and 12.0 μm .

The TOA reflectance of the Earth is computed according to the equation:

$$\rho_{\lambda} = (\pi \cdot L_{\lambda} \cdot d^2) / (ESUN_{\lambda} \cdot \cos \theta_s) \quad (6.13)$$

Where ρ_{λ} = Planetary TOA reflectance [unitless]

π = Mathematical constant equal to ~ 3.14159 [unitless]

L_λ = Spectral radiance at the sensor's aperture [$\text{W} / (\text{m}^2 \text{ sr } \mu\text{m})$]

d = Earth – Sun distance [astronomical units]

ESUN_λ = Mean exoatmospheric solar irradiance [$\text{W} / (\text{m}^2 \mu\text{m})$]

θ_s = Solar zenith angle [degrees]

The Landsat images were processed with necessary calibrations to obtain the TOA reflectance values. The TOA reflectance values at Band 6 (2200 nm) are representative of the surface reflectance and hence, for retrieving the AOD at Band 1 (470 nm) & Band 3 (660 nm) corresponding to their TOA reflectance values, the generated LUTs from radiative transfer approach were considered. Table 6.3 shows the values used to calculate the reflectances using equation (6.13).

Date (Julian day)	Solar Zenith Angle (θ_s in deg)	d (Earth–Sun distance in astronomical units)	ESUN_λ (in $\text{W}/(\text{m}^2 \mu\text{m})$)
021902 (50)	56.94	0.98856	<u>Band 1 (0.450 – 0.515 μm):</u> 1996
081402 (226)	33.52	1.01299	<u>Band 3 (0.633 – 0.690 μm):</u> 1536
092402 (267)	45.18	1.00318	<u>Band 6 (2.080 – 2.350 μm):</u> 82.38

Table 6.3 Constants values used in the calculation of the TOA reflectance for the observation days in our present study. Date is given in mmddyy format with the corresponding Julian day in parentheses.

6.3 Observations and Results:

In our present study, we consider the three pairs of locations with urban and vegetation classes. Each class consists of 10 X 10 boxes of contiguous pixels (300 m x

300m area) selected from three different locations within New York City. In The Bronx at the top (Site 1), Manhattan in the middle (Site 2) and Brooklyn at the bottom (Site 3), as shown in figure 5.1. The large white rectangles contains the two classes urban (blue (gray color in black and white)) and vegetation (yellow (light color in black and white)). Each pair of urban and vegetation regions are within a couple hundred meters from each other.

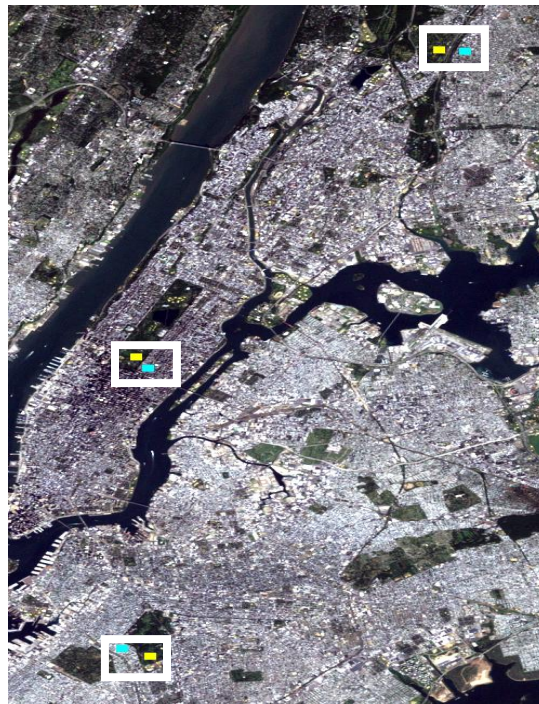


Figure 6.1 Image over New York City from Landsat showing the selected areas of analysis. The boxes surround the locations of the selected sites. Each box encloses separate urban (blue (gray color in black and white)) and vegetation (yellow (light color in black and white)) regions.

In order to check the consistency in the retrieved AOD with those observed instrumentally, we examine the days with low AOD and high AOD observation at CCNY

In order to stabilize the retrieval, we eliminate the high and low surface reflectance resulting AODs from the observations by 20%. The results for the high AOD case are shown next in figures 6.2 (Site 1), 6.3 (Site 2) and 6.4 (Site 3).

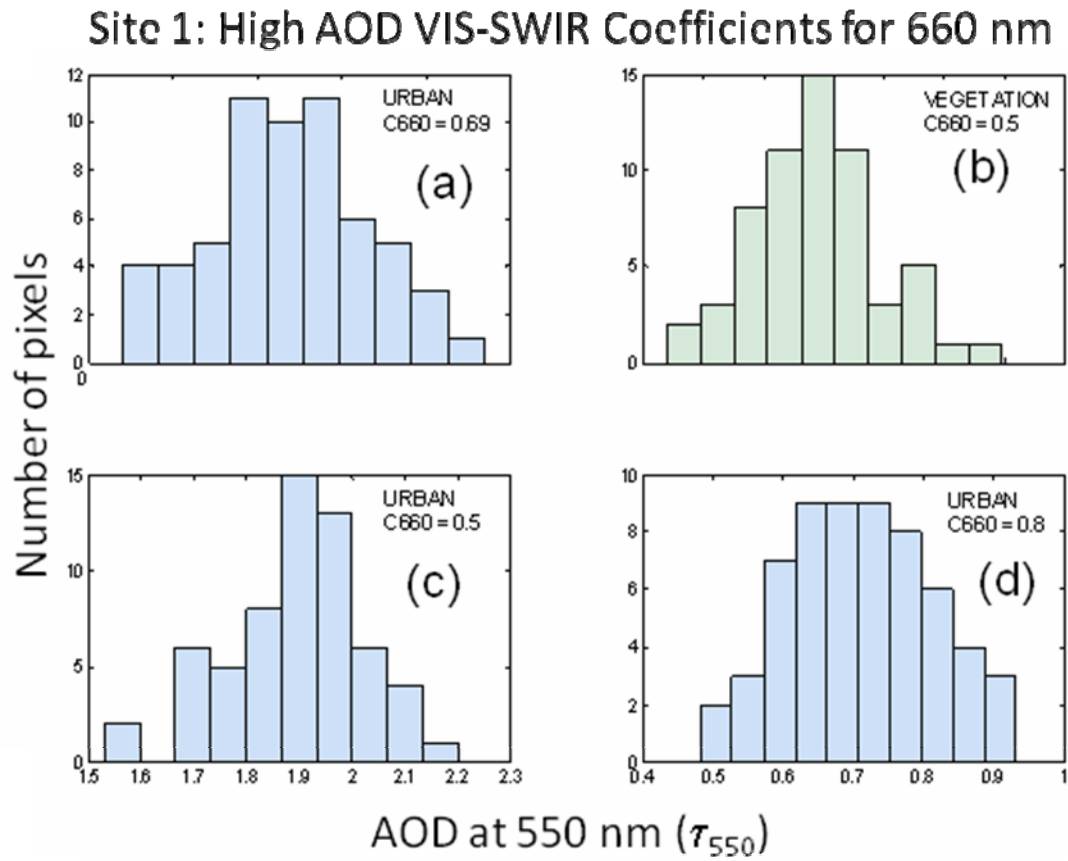


Figure 6.2 Site 1 histogram of 10 x 10 pixels boxes from day with high AOD (August 14, 2002). The x-axis shows the values of AOD at 550 nm, the y-axis shows the number of pixels. Panel (b) shows the vegetation region using $C_{660}=0.5$. Panels (a), (c) and (d) shows the urban region using $C_{660}=0.69$, $C_{660}=0.5$, and $C_{660}=0.8$ respectively.

We can see that the best agreement between the vegetation and urban occurs when the urban coefficient is chosen as $C_{\text{urban}}(660,2130) \sim .69$. If the vegetation land surface model is used to process urban scene ($C(660,2130)=0.5$) there will be a large overestimation as shown on panel (c) in figure 6.2. The dispersion of AOD using 300 meters resolution is $<10\%$. It is quite impressive but it is a function of the high AOD for this day (and using more accurate phase function from AERONET).

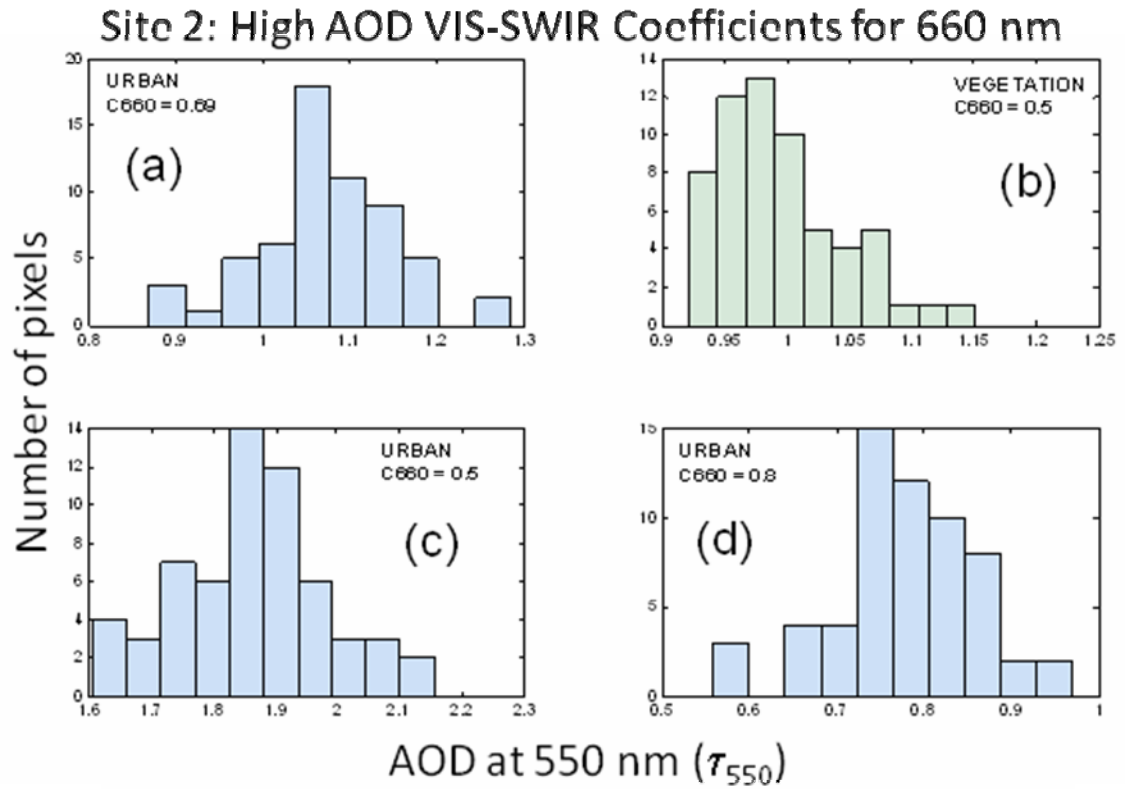


Figure 6.3 Site 2 histogram of 10 x 10 pixels boxes from day with high AOD (August 14, 2002). The x-axis shows the values of AOD at 550 nm, the y-axis shows the number of pixels. Panel (b) shows the vegetation region using $C_{660}=0.5$. Panels (a), (c) and (d) shows the urban region using $C_{660}=0.69$, $C_{660}=0.5$, and $C_{660}=0.8$ respectively.

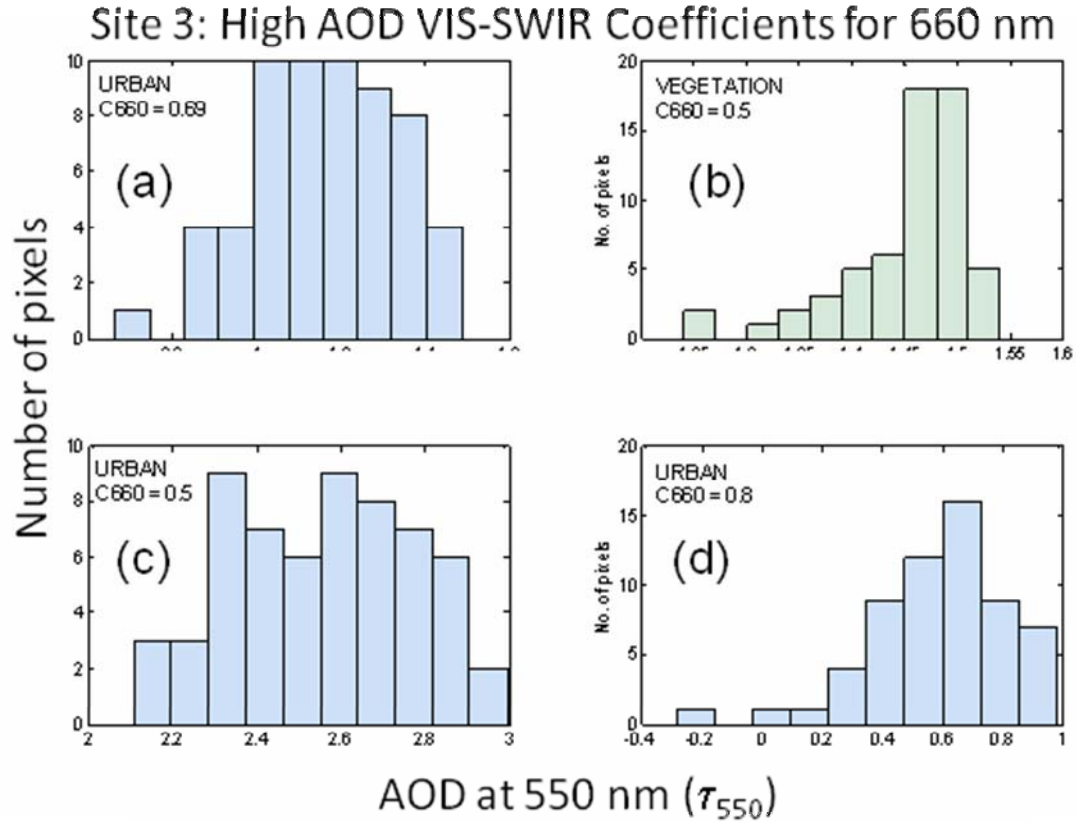


Figure 6.4 Site 3 histogram of 10 x 10 pixels boxes from day with high AOD (August 14, 2002). The x-axis shows the values of AOD at 550 nm, the y-axis shows the number of pixels. Panel (b) shows the vegetation region using C660=0.5. Panels (a), (c) and (d) shows the urban region using C660=0.69, C660=0.5, and C660=0.8 respectively.

We note first of all that the for the high AOD case, AERONET has an AOD at 550 nm of 0.8. The retrievals that we obtain range from .26 (sites 1 and 2) to .42. Clearly, the retrievals are a significant underestimate of the AERONET AOD. However, this might be explained by the fact that in our model, we assume the entire ground reflection in the 2130 band is ground. However, for this day, the AOD at 2130 was about .04. This contribution to the reflection should be made which will result in significantly smaller

visible channel reflectances and correspondingly higher AOD retrieval for both urban and vegetation pixels. On the other hand, it is important to note that the choice of spectral slopes results in the best consistency. In particular, processing the urban with too small (.5) or too large a spectral slope (.8) results in corresponding too high or too low an AOD when compared to the vegetation results. The results for low AOD were very inconsistent as shown in figures 6.5 (Site 1), 6.6 (Site 2) and 6.7 (Site 3).

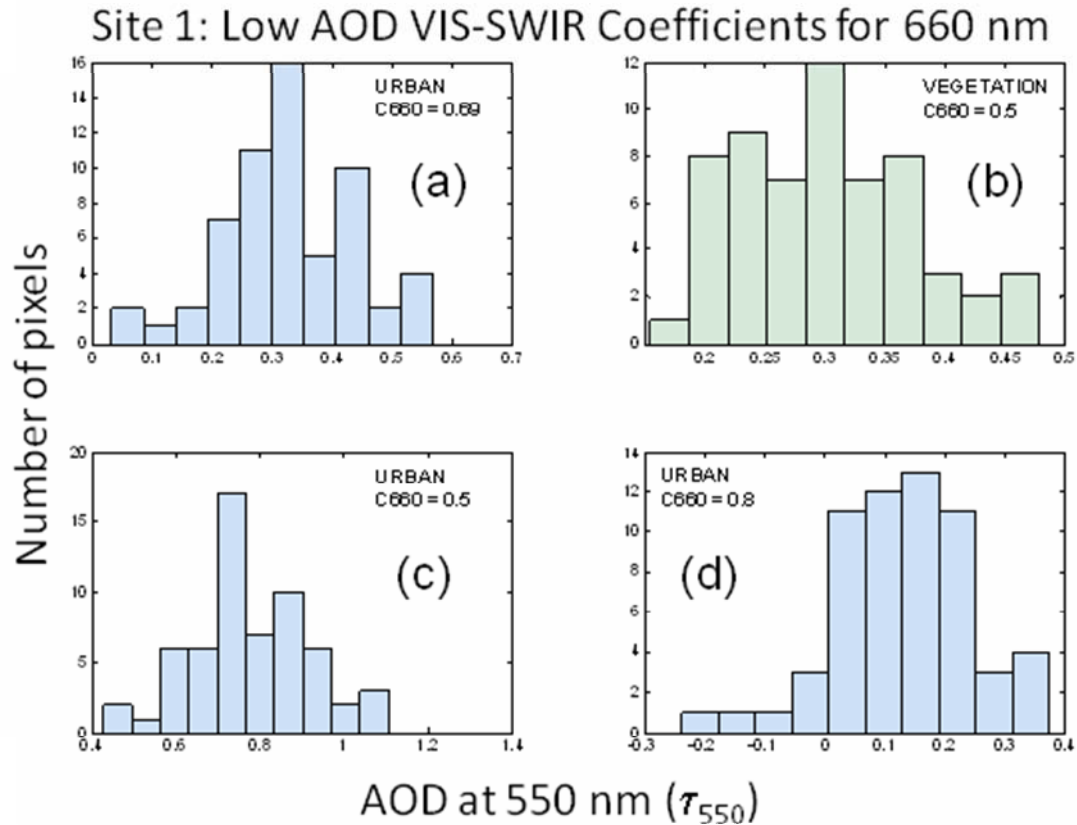


Figure 6.5 Site 1 histogram of 10 x 10 pixels boxes from day with low AOD (February 19, 2002). The x-axis shows the values of AOD at 550 nm, the y-axis shows the number of pixels. Panel (b) shows the vegetation region using $C_{660}=0.5$. Panels

(a), (c) and (d) shows the urban region using $C_{660}=0.69$, $C_{660}=0.5$, and $C_{660}=0.8$ respectively.

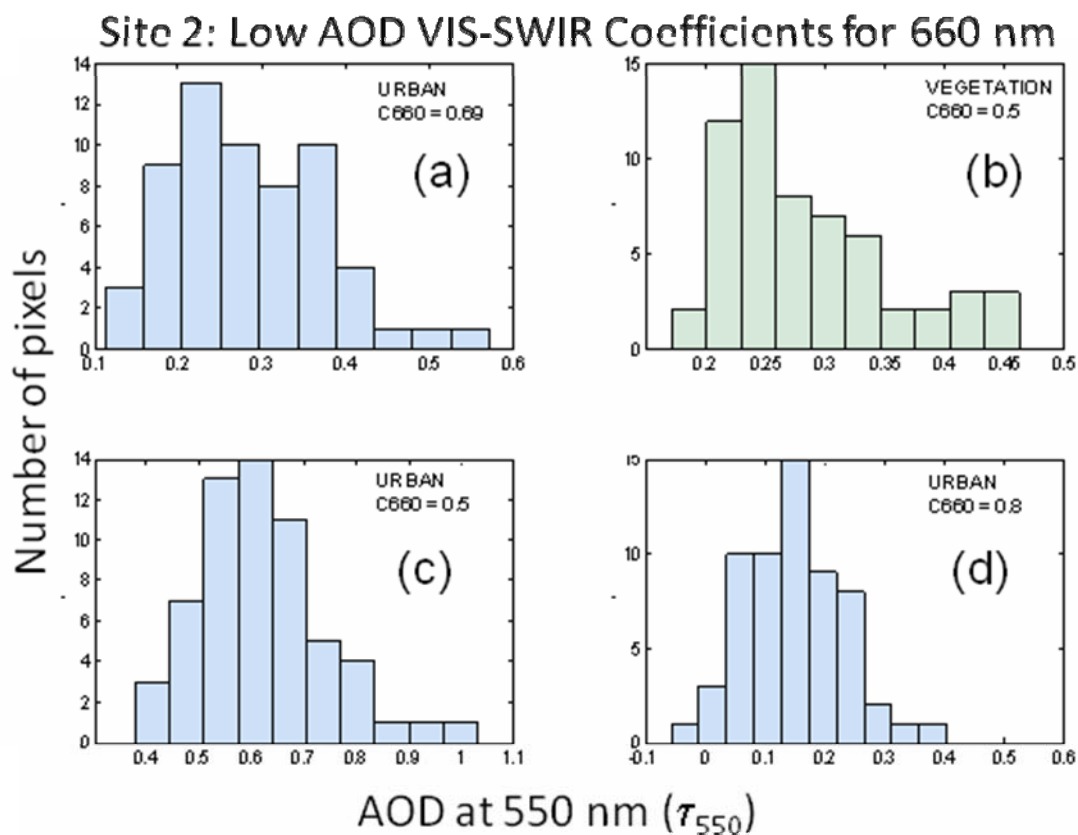


Figure 6.6 Site 2 histogram of 10 x 10 pixels boxes from day with low AOD (February 19, 2002). The x-axis shows the values of AOD at 550 nm, the y-axis shows the number of pixels. Panel (b) shows the vegetation region using $C_{660}=0.5$. Panels (a), (c) and (d) shows the urban region using $C_{660}=0.69$, $C_{660}=0.5$, and $C_{660}=0.8$ respectively.

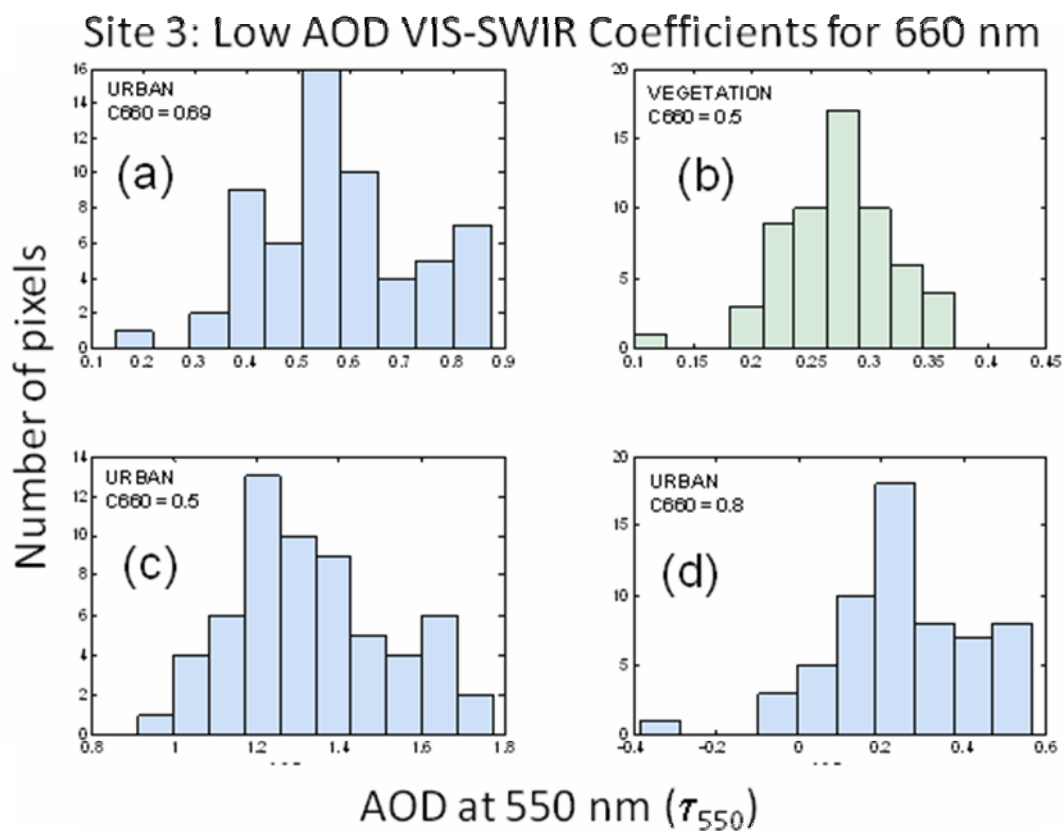


Figure 6.7 Site 3 histogram of 10 x 10 pixels boxes from day with low AOD (February 19, 2002). The x-axis shows the values of AOD at 550 nm, the y-axis shows the number of pixels. Panel (b) shows the vegetation region using C660=0.5. Panels (a), (c) and (d) shows the urban region using C660=0.69, C660=0.5, and C660=0.8 respectively.

Days of observation	Locations	URBAN AOD $\pm$ sd			VEGETATION AOD $\pm$ sd
		C660 urb=0.69	C660 urb=0.8	C660 urb=0.5	C660 veg=0.5
August 14, 2002 High AOD	Site 1	1.019 $\pm$ 0.163	1.896 $\pm$ 0.248	0.674 $\pm$ 0.214	0.997 $\pm$ 0.057
	Site 2	1.062 $\pm$ 0.100	1.852 $\pm$ 0.121	0.767 $\pm$ 0.122	1.0 $\pm$ 0.051
	Site 3	1.008 $\pm$ 0.417	2.214 $\pm$ 0.848	0.514 $\pm$ 0.343	1.416 $\pm$ 0.117
	CCNY AERONET	Day's Mean $\tau_{500} = 0.946$ (0.98 at the time of observation)			
**August 14, 2002 High AOD	Site 1	1.037 $\pm$ 0.096	1.899 $\pm$ 0.131	0.709 $\pm$ 0.105	0.993 $\pm$ 0.053
	Site 2	1.070 $\pm$ 0.082	1.859 $\pm$ 0.121	0.779 $\pm$ 0.081	0.994 $\pm$ 0.049
	Site 3	2.561 $\pm$ 0.223	0.593 $\pm$ 0.233	0.593 $\pm$ 0.233	1.454 $\pm$ 0.060
	CCNY AERONET	Day's Mean $\tau_{500} = 0.946$ (0.98 at the time of observation)			
February 19, 2002 Low AOD	Site 1	0.321 $\pm$ 0.127	0.782 $\pm$ 0.177	0.123 $\pm$ 0.134	0.305 $\pm$ 0.087
	Site 2	0.281 $\pm$ 0.103	0.625 $\pm$ 0.155	0.141 $\pm$ 0.094	0.286 $\pm$ 0.078
	Site 3	0.568 $\pm$ 0.162	1.331 $\pm$ 0.213	0.234 $\pm$ 0.217	0.285 $\pm$ 0.050
	CCNY AERONET	Day's Mean $\tau_{500} = 0.047$ (0.038 at the time of observation)			
**February 19, 2002 Low AOD	Site 1	0.325 $\pm$ 0.113	0.780 $\pm$ 0.140	0.131 $\pm$ 0.115	0.302 $\pm$ 0.074
	Site 2	0.283 $\pm$ 0.092	0.622 $\pm$ 0.123	0.147 $\pm$ 0.084	0.284 $\pm$ 0.071
	Site 3	0.579 $\pm$ 0.153	1.34 $\pm$ 0.195	0.255 $\pm$ 0.175	0.272 $\pm$ 0.048
	CCNY AERONET	Day's Mean $\tau_{500} = 0.047$ (0.038 at the time of observation)			

Table 5.4 Comparative summary of mean AOD observed from the selected grid regions (Sites 1, 2 & 3) for days of observation with low and high AOD as measured by the AERONET site at CCNY. The AOD values (τ_{550}) correspond to those retrieved from the LUTs obtained by considering C660= 0.5, C660 = 0.69 and C660 = 0.8 (where C660 = $\rho_g(660) / \rho_g(2130)$).

With the low AOD from AERONET in this case, we cannot hope to obtain very good accuracy but we do see that the retrievals are consistent with low AOD in general and again the best consistency between the vegetation and urban scenes when the

optimum spectral slopes are used. Note in particular that if the urban scenes are processed using the coefficient values from vegetation, a much higher AOD will result. In addition, we see very good consistencies across the 3 sites, which are expected for clear sky cases where homogeneity is expected. The results of this analysis are provided in table 6.3 where we can observe that there is a large mismatch of observed AOD from CCNY AERONET with the retrieved mean on the day of low mean AOD, while the values look very consistent for high AOD days. Site 3 give very inconsistent results for both low and high AOD days and needs to be further studied.

CHAPTER 7

Conclusions

Measuring aerosols in urban areas is critical in providing useful air quality information to the residents. In particular, high aerosol loadings can lead to adverse health effects resulting in the issuance of bad air quality announcements. While the EPA has a network of PM_{2.5} and PM₁₀ sensors in many major cities like New York City, these sensors are mostly located on small buildings in order to take measurements close to the street to keep track of the aerosols reaching the population. While this is good for providing information about the present levels of aerosols at a particular location, these sensors cannot help to predict the type or amount of aerosols that will be present on the same area in a near future. Thus this network of sensors cannot be used as part of an early warning system of incoming aerosols plumes, which would be greatly helpful for people with a compromised respiratory system. A much better approach for such a system would be the use of the satellite sensors which can currently identify and quantify aerosols.

In general, satellites can provide a very good measurement of aerosols over large bodies of water, and heavily vegetated areas, but there are some inconsistencies with these measurements over urban areas. The algorithms for these sensors need to be modified to improve the accuracy of the measurements of aerosols over large urban areas.

The main point of this thesis is that for urban scenes, the spectral ratios between the VIS and MIR have to be adjusted. The usefulness of this is that these slopes are fairly stable and one obtained should be robust. The optimal values that we got from a variety of methods confirm this. These methods include

1) Combined MODIS and AERONET measurements with AERONET providing us with ancillary atmospheric data which allows the retrieval of surface reflection at both the SWIR and VIS bands directly. The results of this exercise led to values of the VIS-SWIR spectral ratios which were significantly higher than those assumed by the MODIS operational algorithm which would explain the observed overestimates in AOD seen by MODIS. These observed biases in AOD can then be removed when using the refined VIS-SWIR relations as demonstrated by Dr. Min Oo1

2) A Time series approach where the MODIS VIS and SWIR TOA signals are overlaid on Rayleigh scatter signals for the relevant geometries. In particular, we found at least for urban scenes that the minimum signals were consistent with the spectral ratios obtained using method 1.

3) Based on the above methods, it was important to test our hypothesis that the urban pixels have a significantly larger ratio than vegetation. To do this, we used high spatial resolution sensors Hyperion and Landsat and performed atmospheric correction using visibility estimates from AERONET to show that the VIS to SWIR ratios used in MODIS collect 5 over land algorithm are indeed different for vegetation and urban pixels. Quantitative analysis shows that for both urban and vegetation classes, the ratios are

in good agreement with the other two methods. The agreement between the two sensors is particularly impressive since Landsat 7 (ETM+) sensor does not have the narrow bandwidth of the Hyperion sensor.

In summary, when comparing the different methods, the final valuye we take as the best ones are: $c_{2130}^{470} = .38 \pm .08$; $c_{2130}^{660} = .69 \pm .12$

Finally, we also illustrated the need for using the correct spectral ratios when processing Landsat data. Good agreement with AOD in neighboring vegetation and urban scenes was possible only with the updated values, even though observations in winter were often not consistent. For high AOD case, a very good agreement was found between Landsat and CIMEL AOD which was to be expected since we used accurate phase function model from CIMEL. Good agreement was also found between the 470 nm and 660 nm channels, when using either band 1 (470nm) or band 3 (660nm), from Landsat imagery, with the correct AOD. It is expected that using both bands simultaneously would lead to even better improvement.

However, when the AOD from the CIMEL was low, (i.e $< .1$), although the retrievals for both vegetation and urban were consistent, the values were quite larger than those from CIMEL. This mismatch is under investigation. In particular, for Landsat, the use of these coefficients makes the retrieval of AOD over vegetation and urban pixels consistent. Without this change, a severe discontinuity in the AOD retrieval would be seen at vegetation / urban boundaries within the urban canopy.

Future Work

We need to perform analysis over a much larger set of urban areas to see robustness of these approaches using the Landsat imagery. This is now possible since USGS now has opened these images to the public free of charge. We need to match up against more moderate AOD cases to see how accurate we can monitor AOD. We also need to assess spatial resolution to see how small we can make the scenes (size of pixels box) and get reasonable AOD values. Finally, we need to get a better understanding of the correlation ratio for winter scenes.

Appendix A

Landsat README File

U.S. Geological Survey
LEVEL-1 PRODUCT README: GeoTIFF

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INTRODUCTION

This product was created by the U.S. Geological Survey (USGS) and contains Landsat data files in Geographic Tagged Image-File Format (GeoTIFF).

The Landsat 1, 2, and 3 satellites carried the Multispectral Scanner (MSS) sensor; the Landsat 4 and 5 satellites carry both the MSS and the Thematic Mapper (TM) sensors; and the Landsat 7 satellite carries the Enhanced Thematic Mapper Plus (ETM+) sensor. These sensors support the Landsat Project's mission to provide quality remote sensing data in support of research and applications activities.

For information on MSS products (Landsats 1-5):
<http://eros.usgs.gov/products/satellite/mss.php>

For information on TM products (Landsats 4-5):
<http://eros.usgs.gov/products/satellite/tm.php>

For information on ETM+ products (Landsat 7):
<http://eros.usgs.gov/products/satellite/landsat7.php>

For information on the Landsat Project, the Landsat Data Continuity Mission (LDCM), project documentation, and data acquisitions:
<http://landsat.usgs.gov>

For detailed information on the Landsat 7 satellite, the ETM+ sensor, and Landsat 7 data products:

<http://landsathandbook.gsfc.nasa.gov/handbook.html>

FORMAT

Landsat MSS, TM, and ETM+ data are provided in GeoTIFF for Level 1T (terrain corrected) products, or for either Level 1Gt (systematic terrain corrected) or Level 1G (systematic corrected) products, if Level 1T processing is not available. GeoTIFF defines a set of publicly available TIFF tags that describe cartographic and geodetic information associated with TIFF images. GeoTIFF is a format that enables referencing a raster image to a known geodetic model or map projection.

The initial tags are followed by image data that, in turn, may be interrupted by more descriptive tags. By using the GeoTIFF format, both metadata and image data can be encoded into the same file.

The Landsat 7 ETM+ GeoTIFF file format is described in detail in the "Landsat 7 ETM+ Level 1 Product Data Format Control Book (DFCB), LS-DFCB-04":

http://landsat.usgs.gov/documents/LS_DFCB_04.pdf

The Landsat 4-5 TM GeoTIFF file format is described in detail in the "Landsat Thematic Mapper (TM) Level 1 (L1) Data Format Control Book (DFCB), LS-DFCB-20":

http://landsat.usgs.gov/documents/LS_DFCB_20_TM_Level_1_DFCB.pdf

For more information on GeoTIFF visit:

<http://trac.osgeo.org/geotiff>

ORGANIZATION

Each band of Landsat data in the GeoTIFF format is delivered as a grayscale, uncompressed, 8-bit string of unsigned integers. A metadata (MTL) file is included with data processed through the Level-1 Product Generation System (LPGS). A file containing the ground control points (GCP) used during image processing is also included. A processing history (WO) file is included with data processed through the National Landsat Archive Production System (NLAPS). Landsat 7 ETM+ SLC-off products processed after December 11, 2008, will include an additional directory (gap_mask) that contains a set of flat binary scan gap mask files (one per band). (Please note that the processing date and acquisition date are not necessarily the same.)

* DATA FILE NAMES

The file naming convention for Landsat LPGS-processed GeoTIFF data is as follows:

LLfpppprrr_rrrYYYYMMDD_AAA.TIF where:

LL	= Landsat sensor (LE for ETM+ data; LT for TM data)
f	= ETM+ data format (1, 2, or G) (character omitted from TM file name)
ppp	= starting path of the product

```

rrr_rrr      = starting and ending rows of the product
YYYYMMDD    = acquisition date of the image
AAA          = file type:
    B10      = band 1
    B20      = band 2
    B30      = band 3
    B40      = band 4
    B50      = band 5
    B61      = band 6L (low gain)
    B62      = band 6H (high gain)
    B70      = band 7
    B80      = band 8
    MTL      = Level-1 metadata
    GCP      = ground control points
TIF          = GeoTIFF file extension

```

The file naming convention for Landsat NLAPS-processed GeoTIFF data is as follows:

```

LLNpppprrrOOYYDDMM_AA.TIF  where:
LL      = Landsat sensor (LM for MSS data; LT for TM data)
N       = satellite number
ppp     = starting path of the product
rrr     = starting row of the product
OO      = WRS row offset (set to 00)
YY      = last two digits of the year of
          acquisition
DDD     = Julian date of acquisition
MM      = instrument mode (10 for MSS; 50 for TM)
AA      = file type:
    B1    = band 1
    B2    = band 2
    B3    = band 3
    B4    = band 4
    B5    = band 5
    B6    = band 6
    B7    = band 7
    W0    = processing history file
TIF      = GeoTIFF file extension

```

* GAP MASKS

All Landsat 7 ETM+ SLC-off imagery processed on or after December 11, 2008, will include gap mask files. (Please note the difference between acquisition date and processing date.) The gap mask files are bit mask files showing the locations of the image gaps (areas that fall between ETM+ scans). One tarred and gzip-compressed gap mask file is provided for each band in GeoTIFF format. The file naming convention for gap mask files is identical to that described above for LPGA-processed GeoTIFF data, with the third character (data format) set as 'G'.

If gap mask files are not included with the data, a tutorial for creating them can be found at:
http://landsat.usgs.gov/gap_mask_files_are_not_provided_can_I_create_my_own.php

* README

The README_GTF.TXT (or README.GTF) is an ASCII text file and is this file.

* READING DATA

Delivered via file transfer protocol (FTP): data files are tarred and g-zip compressed and will need to be unzipped and untarred before the data files can be used. UNIX systems should have the "gunzip" and "tar" commands available for uncompressing and accessing the data. For PC users, free software can be downloaded from an online source.

Otherwise, check your PC, as you may already have appropriate software available.

No software is included on this product for viewing Landsat data.

GENERAL INFORMATION and DOCUMENTATION

Landsat and the Landsat Data Continuity Mission (LDCM):
<http://landsat.usgs.gov>

Landsat 7 (ETM+):
<http://eros.usgs.gov/products/satellite/landsat7.php>

Landsat 4-5 (TM):
<http://eros.usgs.gov/products/satellite/tm.php>

Landsat 1-5 (MSS):
<http://eros.usgs.gov/products/satellite/mss.php>

Landsat data access:

- * USGS Global Visualization Viewer (GloVis): <http://glovis.usgs.gov>
- * USGS EarthExplorer: <http://earthexplorer.usgs.gov>
- * Landsat International Ground Station (IGS) network:
http://landsat.usgs.gov/about_ground_stations.php

FGDC metadata:

<http://clearinghouse1.fgdc.gov/fgdc/EDCgateway.html>

Data restrictions and citation:

<http://eros.usgs.gov/about/customer/citation.html>

* National Snow and Ice Data Center (NSIDC)
 Radarsat Antarctic Mapping Project (RAMP) elevation data citation:

Liu, H., K. Jezek, B. Li, and Z. Zhao. 2001.
 Radarsat Antarctic Mapping Project digital elevation model version 2.
 Boulder, CO: National Snow and Ice Data Center. Digital media.

For information on the data, please refer to the data set documentation available at the following web site:

<http://nsidc.org/data/nsidc-0082.html>

PRODUCT SUPPORT

For further information on this product, contact USGS
EROS Customer Services:

Customer Services (ATTN: Landsat)
U.S. Geological Survey
Earth Resources Observation and Science (EROS) Center
47914 252nd Street
Sioux Falls, SD 57198-0001

Tel: 800-252-4547

Tel: 605-594-6151

Fax: 605-594-6589

Email: custserv@usgs.gov

Web: <http://eros.usgs.gov/about/customer/index.html>

For information on other products from USGS EROS:
<http://eros.usgs.gov/>

For information on other USGS products:
<http://ask.usgs.gov/>
or call 1-888-ASK-USGS (275-8747)

DISCLAIMER

Any use of trade, product, or firm names is for descriptive
purposes only and does not imply endorsement by the U.S.
Government.

Publication Date: February 2009

Appendix B

Landsat Metadata File

```

GROUP = L1_METADATA_FILE
  GROUP = METADATA_FILE_INFO
    ORIGIN = "Image courtesy of the U.S. Geological Survey"
    REQUEST_ID = "0800907160153_00020"
    PRODUCT_CREATION_TIME = 2009-07-17T01:42:55Z
    STATION_ID = "EDC"
    LANDSAT4_XBAND = "0"
    LANDSAT7_XBAND = "1"
    GROUND_STATION = "EDC"
    LPS_PROCESSOR_NUMBER = 8
    DATE_HOUR_CONTACT_PERIOD = "0205015"
    SUBINTERVAL_NUMBER = "01"
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  GROUP = PRODUCT_METADATA
    PRODUCT_TYPE = "L1T"
    ELEVATION_SOURCE = "GLS2000"
    PROCESSING_SOFTWARE = "LPGS_10.1.0"
    EPHEMERIS_TYPE = "DEFINITIVE"
    SPACECRAFT_ID = "Landsat7"
    SENSOR_ID = "ETM+"
    SENSOR_MODE = "SAM"
    ACQUISITION_DATE = 2002-02-19
    SCENE_CENTER_SCAN_TIME = 15:28:41.0627195Z
    WRS_PATH = 14
    STARTING_ROW = 32
    ENDING_ROW = 32
    BAND_COMBINATION = "123456678"
    PRODUCT_UL_CORNER_LAT = 41.2971944
    PRODUCT_UL_CORNER_LON = -76.3281797
    PRODUCT_UR_CORNER_LAT = 41.2933963
    PRODUCT_UR_CORNER_LON = -73.3757113
    PRODUCT_LL_CORNER_LAT = 39.3300752
    PRODUCT_LL_CORNER_LON = -76.2901812
    PRODUCT_LR_CORNER_LAT = 39.3265310
    PRODUCT_LR_CORNER_LON = -73.4221733
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    PRODUCT_UL_CORNER_MAPY = 4572600.000
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    PRODUCT_UR_CORNER_MAPY = 4572600.000
    PRODUCT_LL_CORNER_MAPX = 388800.000
    PRODUCT_LL_CORNER_MAPY = 4354200.000
    PRODUCT_LR_CORNER_MAPX = 636000.000
    PRODUCT_LR_CORNER_MAPY = 4354200.000
    PRODUCT_SAMPLES_PAN = 16481
    PRODUCT_LINES_PAN = 14561
    PRODUCT_SAMPLES_REF = 8241
    PRODUCT_LINES_REF = 7281
    PRODUCT_SAMPLES_THM = 4121
  
```

```

PRODUCT_LINES_THM = 3641
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BAND2_FILE_NAME = "L71014032_03220020219_B20.TIF"
BAND3_FILE_NAME = "L71014032_03220020219_B30.TIF"
BAND4_FILE_NAME = "L71014032_03220020219_B40.TIF"
BAND5_FILE_NAME = "L71014032_03220020219_B50.TIF"
BAND61_FILE_NAME = "L71014032_03220020219_B61.TIF"
BAND62_FILE_NAME = "L72014032_03220020219_B62.TIF"
BAND7_FILE_NAME = "L72014032_03220020219_B70.TIF"
BAND8_FILE_NAME = "L72014032_03220020219_B80.TIF"
GCP_FILE_NAME = "L71014032_03220020219_GCP.txt"
METADATA_L1_FILE_NAME = "L71014032_03220020219_MTL.txt"
CPF_FILE_NAME = "L7CPF20020101_20020331_06"
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  LMIN_BAND1 = -6.200
  LMAX_BAND2 = 196.500
  LMIN_BAND2 = -6.400
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  LMAX_BAND4 = 157.400
  LMIN_BAND4 = -5.100
  LMAX_BAND5 = 31.060
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  LMIN_BAND7 = -0.350
  LMAX_BAND8 = 243.100
  LMIN_BAND8 = -4.700
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GROUP = MIN_MAX_PIXEL_VALUE
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CORRECTION_METHOD_GAIN_BAND4 = "CPF"
CORRECTION_METHOD_GAIN_BAND5 = "CPF"
CORRECTION_METHOD_GAIN_BAND61 = "CPF"
CORRECTION_METHOD_GAIN_BAND62 = "CPF"
CORRECTION_METHOD_GAIN_BAND7 = "CPF"
CORRECTION_METHOD_GAIN_BAND8 = "CPF"
CORRECTION_METHOD_BIAS = "IC"
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BAND2_GAIN = "H"
BAND3_GAIN = "H"
BAND4_GAIN = "H"
BAND5_GAIN = "H"
BAND6_GAIN1 = "L"
BAND6_GAIN2 = "H"
BAND7_GAIN = "H"
BAND8_GAIN = "L"
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BAND2_GAIN_CHANGE = "0"
BAND3_GAIN_CHANGE = "0"
BAND4_GAIN_CHANGE = "0"
BAND5_GAIN_CHANGE = "0"
BAND6_GAIN_CHANGE1 = "0"
BAND6_GAIN_CHANGE2 = "0"
BAND7_GAIN_CHANGE = "0"
BAND8_GAIN_CHANGE = "0"
BAND1_SL_GAIN_CHANGE = 0
BAND2_SL_GAIN_CHANGE = 0
BAND3_SL_GAIN_CHANGE = 0
BAND4_SL_GAIN_CHANGE = 0
BAND5_SL_GAIN_CHANGE = 0
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BAND6_SL_GAIN_CHANGE2 = 0
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BAND8_SL_GAIN_CHANGE = 0
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SUN_ELEVATION = 33.0561950
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END_GROUP = PRODUCT_PARAMETERS
GROUP = CORRECTIONS_APPLIED
STRIPING_BAND1 = "NONE"
STRIPING_BAND2 = "NONE"
STRIPING_BAND3 = "NONE"
STRIPING_BAND4 = "NONE"
STRIPING_BAND5 = "NONE"
STRIPING_BAND61 = "NONE"
STRIPING_BAND62 = "NONE"
STRIPING_BAND7 = "NONE"
STRIPING_BAND8 = "NONE"
BANDING = "N"
COHERENT_NOISE = "Y"
MEMORY_EFFECT = "N"
SCAN_CORRELATED_SHIFT = "N"
INOPERABLE_DETECTORS = "N"
DROPPED_LINES = "N"
END_GROUP = CORRECTIONS_APPLIED
GROUP = PROJECTION_PARAMETERS

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REFERENCE_ELLIPSOID = "WGS84"  
GRID_CELL_SIZE_PAN = 15.000  
GRID_CELL_SIZE_THM = 60.000  
GRID_CELL_SIZE_REF = 30.000  
ORIENTATION = "NUP"  
RESAMPLING_OPTION = "CC"  
MAP_PROJECTION = "UTM"  
END_GROUP = PROJECTION_PARAMETERS  
GROUP = UTM_PARAMETERS  
    ZONE_NUMBER = 18  
END_GROUP = UTM_PARAMETERS  
END_GROUP = L1_METADATA_FILE  
END
```

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PUBLICATIONS AND CONTRIBUTIONS

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2. Min M. Oo, Eduardo Hernandez, Matthias Jerg, Barry M. Gross, Fred Moshary and Samir A. Ahmed "High-resolution surface correlation maps for improved resolution and retrieval of aerosols over urban scenes", SPIE EURO Remote Sensing (ERS08 Symposium), 2008
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4. Min M. Oo, Eduardo Hernandez, Matthias Jerg, Barry M. Gross ,Fred Moshary and Samir A. Ahmed, "Improved MODIS Aerosol Retrieval Using Modified VIS/MIR Surface Albedo Ratio Over Urban Scenes" CoRP Science Symposium, August 12 - 13, 2008,Corvallis, OR
5. Min M. Oo, Eduardo Hernandez, Matthias Jerg, Barry M. Gross, Fred Moshary, and Samir A. Ahmed," Improved MODIS aerosol retrieval over urban scenes" Processing of. SPIE 6684, 66840Q (2007)
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7. Selected as a NASA Student Ambassador, 2009.